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STRUCTURAL ANALYSIS  
OF THE LOWER PALEOZOIC  
ROCKS OF WESTERN GASPE, QUEBEC

by

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A thesis

submitted in partial fulfillment of the requirements

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in the

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A B S T R A C T

The Shickshock Group of Cambrian-Ordovician metavolcanic and metasedimentary rocks outcrops along mountains of the same name bounded to the south by the Shickshock Fault. Within these rocks a schistosity  $S_1$  and layering are parallel, and there is a lineation  $L_1$  marked by a mineral elongation and fine striation. These structures suggest the presence of isoclinal  $F_1$  folds, and the Shickshock rocks are considered to form the core of a major recumbent  $F_1$  fold. The absence of  $F_1$  structures from Lower Ordovician rocks to the north suggests a pre-Lower Ordovician period of folding.

$F_2$  folds, the dominant structures in the area, deform all Paleozoic rocks older than upper Middle Devonian and form the stratigraphically defined Gaspe Synclinorium, an Acadian structure.

On the basis of the inclination and symmetry of the small scale  $F_2$  folds, the Gaspe synclinorium is divided into three major domains.

- I. The northern limb of the synclinorium, north of the Shickshock Fault, where folds in Cambrian to Ordovician rocks are overturned northwest; anticlines have short northwest limbs.
- II. The northern limb, south of the fault to the axial area of the synclinorium, where folds in Siluro-Devonian rocks are upright; folds are both asymmetric as in domain I and symmetric.
- III. The southern limb of the synclinorium where axial planes of folds in Ordovician to Devonian rocks are inclined northwest; anticlines have short southeast limbs.

(ii)

The folds in domains I and II are gradational in inclination and symmetry, but are separated by vertical displacement across the Shickshock Fault.

In the study area, the  $F_2$  folds affect the  $S_1$  schistosity and trend northeast, co-axially with  $F_1$  structures. However, to the northeast, near Mount Albert,  $F_1$  and  $F_2$  structures are divergent.

The  $F_2$  folds vary in shape with lithology, from open in competent, to isoclinal in incompetent rocks. They are dominantly flattened concentric in competent limestone and sandstone, and similar in style in argillite. The folds are doubly plunging, and the axes of variable plunge lie on planar axial surfaces. Slickensides on axial-plane calcite veins dominantly plunge down the dip. This is regarded as the movement direction during the  $F_2$  folding. Variation in the amount of movement resulted in the curved fold axes. It is suggested that the folds progressed from an initial concentric mechanism determined by the competent beds to a flattening and extension along the axial planes determined by flow of the incompetent beds.

An  $S_2$  cleavage, a close fracture set parallel to the  $F_2$  axial planes, is present in all incompetent rocks except those of the Shickshock Group foliated by  $S_1$ . A classification of the cleavage is suggested by means of the ratio of slip across individual fractures and the average fracture spacing. The cleavage is termed fracture if the ratio is less than .25, and fracture slip if it is .25 or more.

$F_3$  folds, restricted to the northern coastal sections are mostly S-shaped in plan, and plunge steeply to the southeast. A rough  $S_3$  fracture lies parallel to the steep axial planes striking E-W. The folds may have formed by N-S compression and sinistral shear along the northeast striking  $F_2$  structures.

$F_4$  folds, restricted to the belt of Cambrian Ordovician rocks with  $S_1$  schistosity, are mesoscopic to microscopic, open with sharp rectilinear hinges, and resemble kink bands. The folds trending NW-SE are S and Z shaped and are considered as a conjugate system developed in response to a northeast-southwest compression parallel to the general strike of the country rock.

A low plunging fine kink lineation and larger kink bands with low dipping axial planes that deform  $S_2$  occur along the axial area of the synclinorium. The kink bands consistently displace the

(iii)

upper rocks away from the axis. They are considered as late stage structures.

Joint patterns in competent rocks (sandstone, siltstone and limestone) consist of dip, strike, and oblique joints which are normal to bedding. In slates there is a predominance of dip sets accompanied by oblique sets, both of which are steeply inclined. Strike joints in slates are rare and form normal to cleavage. The steep dip joints in slates bear no consistent angular relationship to meso- $F_2$  fold axes. Thus it is suggested that joint planes in folded rocks are independent of attitudes of local folds and are post-kinematic in origin.

VOLUME I

STRUCTURAL ANALYSIS  
OF THE LOWER PALEOZOIC  
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TEXT

C O N T E N T SPage

## I INTRODUCTION

1. Nature of Outcrops and Method of Study	1
2. History of Previous Work	2
History of the Interpretation of Structure of the Shickshock Mountains	5
3. Purpose of the Study	7
4. Acknowledgements	8

## II GEOLOGICAL BACKGROUND

1. Stratigraphy	9
2. Taconic Problem	12
Klippen Theory	12
Stratigraphic Position of Ordovician-Silurian Unconformities and Orogenies	14
Radiogenic Dating of the Ordovician Deformations	18
Summary	20
3. Major Structural Features	21
Gaspé Synclinorium	21
Unconformities	23
Regional Metamorphism	23

## III STRUCTURAL GEOLOGY

1. $F_1$ folds	29
Introduction	29
Profile	29
Planar and Linear Elements	29
Age of the $F_1$ Folds	31
2. $F_2$ Folds	34
Introduction	34
Profile	34
Geometrical Elements	39
Fold Axes	40
Axial Planes	40

	<u>Page</u>
Axial Plane Cleavage	41
Discussion on the Definition of Fracture Cleavage	42
S <sub>2</sub> Cleavage, Spacing and Degree of Folding	43
Relationship of the S <sub>2</sub> Spacing to Lithology	43
Deformation between Fractures	45
Classification of the S <sub>2</sub> Fracture Cleavage	46
Calcite and Quartz Veins	47
Calcite Concentration parallel to the F <sub>2</sub> Axial Planes	49
Slickensides on Calcite Veins	49
Bedding Foliation	51
Regional Variation in Style of F <sub>2</sub> Structures	51
S <sub>2</sub> Cleavage in Different Rock Formations	51
Regional Variation in Attitude of F <sub>2</sub> Axial Planes and S <sub>2</sub> Cleavage	53
Division of the Area into Major Domains with respect to F <sub>2</sub> Fold Profiles and Attitudes	54
Outcrop Pattern in the Domains	55
3. F <sub>3</sub> Folds	56
Introduction and Distribution	56
Profile	56
Planar and Linear Elements	56
4. F <sub>4</sub> Folds	58
Introduction and Distribution	58
Profile	58
Planar and Linear Elements	59
Age of the F <sub>4</sub> Folds	60
5. Late Stage (?) Kinks and Associated Structures	61
Low Dipping <del>Kink Bands</del>	61
Low Dipping Slip Planes and Micro-Kinks	62
6. Joints	63
Introduction	63
Theoretical Background	63
Joint Pattern in Lower Paleozoic Rocks	65
Cambrian-Ordovician Rocks	65
Shickshock Group	65

Cambrian-Ordovician Slates	66
Ordovician Rocks	66
Impure Grey Sandstone	66
Calcareous Slates	67
Limestone and Quartzite	67
Siluro-Devonian Rocks	68
St. Leon Formation	68
Cape Bon Ami Formation	68
Grande Greve Formation	68
York River-Heppel Formations	68
Fortin Slates	69
Ste. Marguerite Volcanics	69
Joint Pattern and $F_2$ Fold Plunge	69
Conclusions	69
7. Faults	71
The Shickshock Fault	71
The Ste. Florence Fault	73
8. Structure of the Shickshock Mountains	74

#### IV CONCLUSIONS

1. Summary and Interpretation	76
2. Tectonic History of the Area	84
3. Comparison of Tectonic History with other Areas in Canadian Appalachians	86

V SUGGESTION FOR FUTURE WORK	89
------------------------------	----

VI REFERENCES	90
---------------	----

<u>VII APPENDICES</u>	97
-----------------------	----

I. STRATIGRAPHY AND LITHOLOGICAL DESCRIPTION OF ROCK TYPES IN THE AREA	97
II. PETROLOGY OF THE IMPURE GREY SANDSTONE	125
III. TRAVERSES NORTH AND NORTHEAST OF MOUNT ALBERT	128
IV. CONODONT ANALYSIS	133
V. GRAPTOLITE FAUNA OF THE MIDDLE ORDOVICIAN SLATE WEDGE IN THE SHICKSHOCK FAULT ZONE	134

LIST OF FIGURES

	Opposite Page
1. Geographic location of the area.	1
2. Structural localities of the area.	2
3. Structural localities of Mount Albert Area.	2
4. Areas previously investigated.	2
5. Previous structural interpretations of the Shickshock Mountains.	5
6. Geological map of western Gaspé area (in pocket).	
6a. Geological cross section.                   "   "	
6b. Gaspé Cleavage Fan.	12
7. Stratigraphic columns in Gaspé region.	17
8. Metamorphic isograds in Mount Logan area.	24
9. $F_1$ structural localities in the area.	29
10. Thickness/wave length ratio in $F_2$ microscopic folds.	35
11. Theoretical thickness/ wave length ratio in a concentric fold.	35
12. Spacing of $S_2$ cleavage/ $\alpha(S_s \wedge S_2)$ curve.	43
13. Diagrammatic features of cleavage fracture 'zone'.	46
14. Orientation of $S_2$ cleavage in the area.	53
15. $F_2$ axial plane inclinations contoured in the area.	53
16. S and Z shapes of $F_2$ folds in the area.	55
17. $F_3$ localities in the area.	56
18. Outcrop patterns of $F_3$ folds.	56
19. S and Z shapes of $F_4$ folds in relation to pitch on $S_1$ surface.	58
20. Location and orientation of late stage kink planes.	61
21. Joint localities in the area.	63
22. Stereographic representation of possible joints in an anticline.	64
23. Diagrammatic joint pattern in competent and incompetent rocks in relation to regional folds.	69
24. Structural interpretation of the Shickshock Mountain area.	74
25. Growth of $F_2$ folds.	78
26. Possible mechanism of formation of $F_3$ folds.	81
27. Compression in each tectonic phase in relation to fold trends.	84

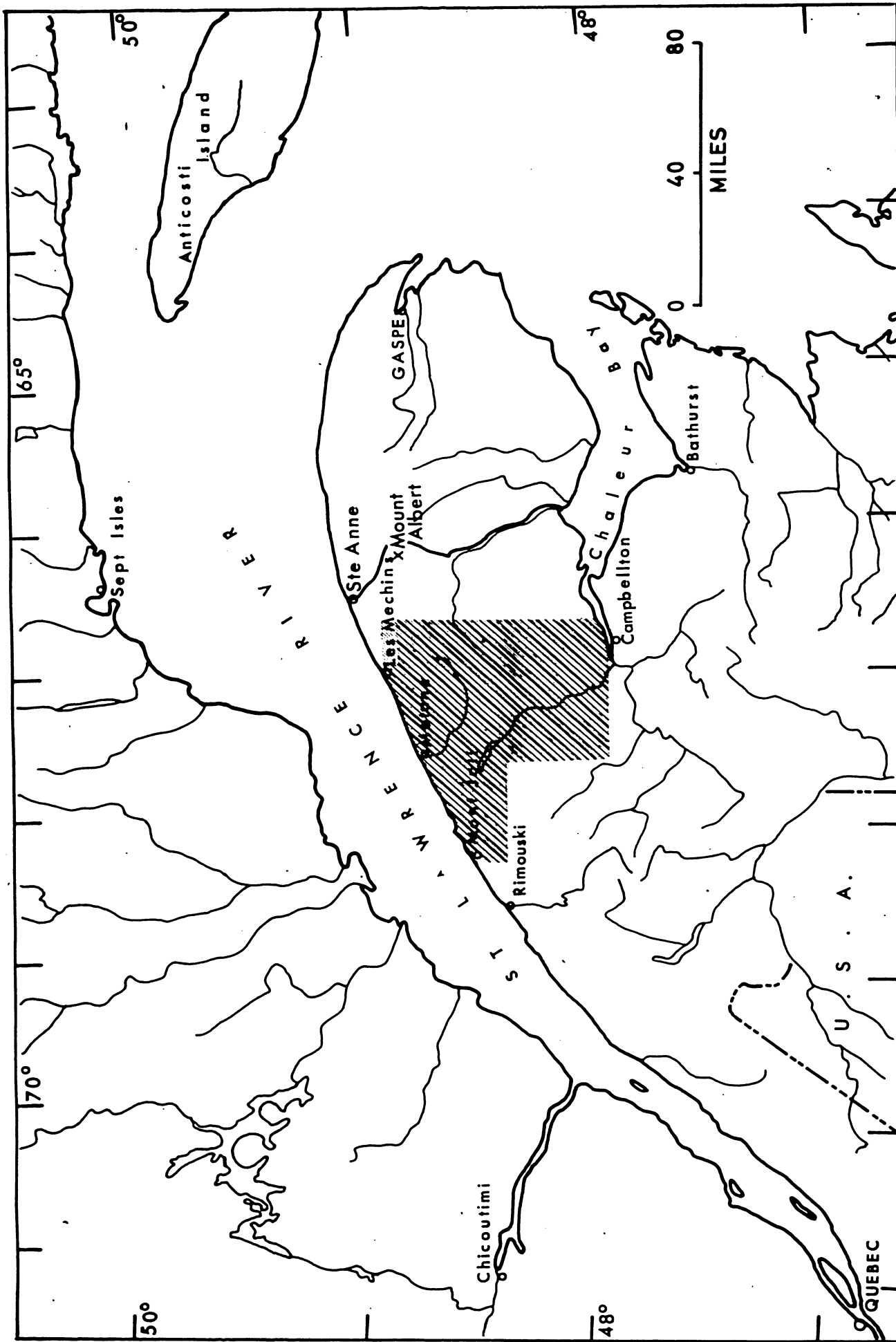
28. Percentage of volcanic fragments in impure grey sandstone.	Opposite Page 127
29. Geological map of the area around Mount Albert.	128

LIST OF TABLES

	Page
1. Table of rock formations.	9
2. Position of unconformities in Cambrian and Ordovician in northern Appalachians.	13
3. Character of fold generations.	26
4. Classification of $S_2$ axial plane cleavage.	47
5. Nature of $S_2$ axial plane cleavage in different rock formations.	52
6. Division of the area into domains according to fold limb ratio and axial plane attitude.	54
7. Percentage composition of impure grey sandstone, Ste. Felicite area.	126

LIST OF PLATES

- I. Net projections of  $F_1$  structures in Cambrian-Ordovician rocks.
- II. a, b, c, d, e.  $F_2$  profiles and variation in bed thickness.
- III. a, b, c, d, e, f, g. Net projections of  $F_2$  structures in Lower Paleozoic rocks.
- IV. Net projections of  $F_2$  structures in Lower Paleozoic rocks to demonstrate geometrical relationships:
  - a. Plotted for localities to demonstrate coincidence of mean  $S_2$  cleavage and  $F_2$  axial plane.
  - b. Relationship between  $L_2$  and  $F_2$  in subareas.
  - c. Upper Ordovician and Siluro-Devonian rocks.
- V. Net projections of slickensides on calcite veins.
- VI. Net projections of  $F_4$  structures in Cambrian-Ordovician rocks.
- VII. Net projections of:
  - a. Joint pattern in Cambrian-Ordovician and Ordovician rocks.
  - b. Joint pattern in Siluro-Devonian rocks.



**FIGURE 1** MAP SHOWING GEOGRAPHIC LOCATION OF STUDY AREA

## I INTRODUCTION

1. Nature of Outcrops and Method of Study

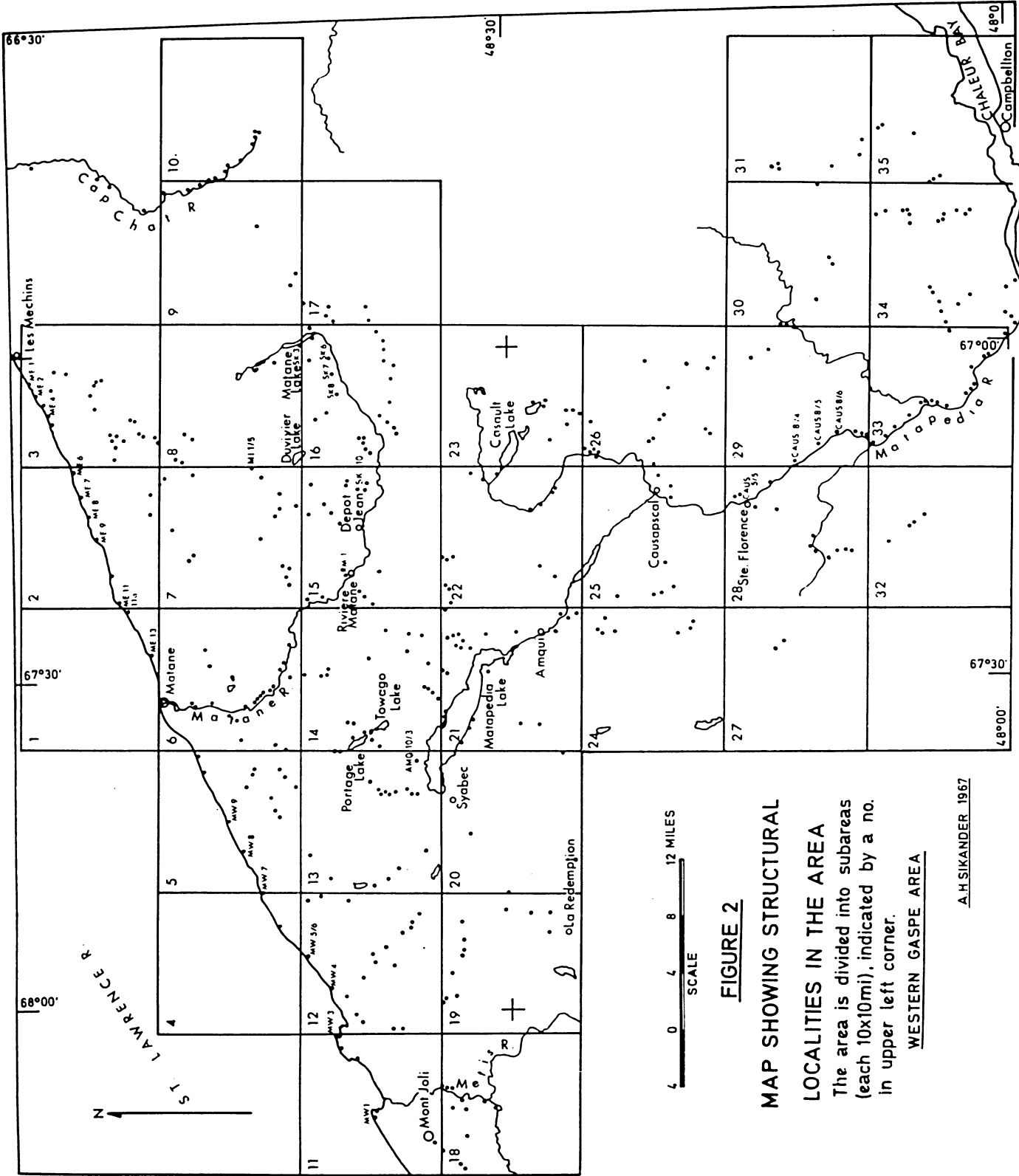
The study area, lying approximately between latitudes 48°N-49°N and longitudes 66°30' W-68°05' W covers the western Gaspé peninsula on both sides of the Matapédia Valley, between the St. Lawrence River and Chaleur Bay (fig. 1). It includes sixty-five miles of well exposed coast along the St. Lawrence River, between Les Méchins and Mont Joli.

Rocks are poorly exposed in the interior owing to thick vegetation. Roadside exposures were found useful inland, especially along Highway 6, which cuts across the general strike of the country rock along the Matapédia River. Fortunately the field work coincided with the country road building programme of the Government of Quebec and hence fresh rock surfaces were frequently available for study. Natural exposures and stream sections were generally found subject to slumping and solifluxion. Outcrops were rejected for structural measurements on the slightest suggestion of the rocks being subject to mass wasting. In the Shickshock Mountains accessible outcrops are extremely scarce owing to dense, practically impenetrable vegetation and high relief.

The available outcrops were investigated in detail by traversing along roads, lumber trails, streams, rivers and the coast. The coastal sections during low tides and freshly cut road sections provided by far the best exposures for the study.

The work was initiated on the northern coastal sections in Ordovician slates, sandstones and limestones for structural details, especially to establish the geometrical relationship between different structural elements. It was then extended inland and southward to Chaleur Bay.

The definitions of rock lithologies as established by previous workers have been accepted in this work. No major synthesised work on the geology of western Gaspé as a whole was available. A synthesis of the region has been made from the reports of a number of workers, especially from those of the Government of Quebec; and their observations have frequently been used in addition to those made by the worker himself.



**FIGURE 2**

**MAP SHOWING STRUCTURAL LOCALITIES IN THE AREA**

The area is divided into subareas (each 10x10mi), indicated by a no. in upper left corner.

**WESTERN GASPE AREA**

A.H.SIKANDER 1967

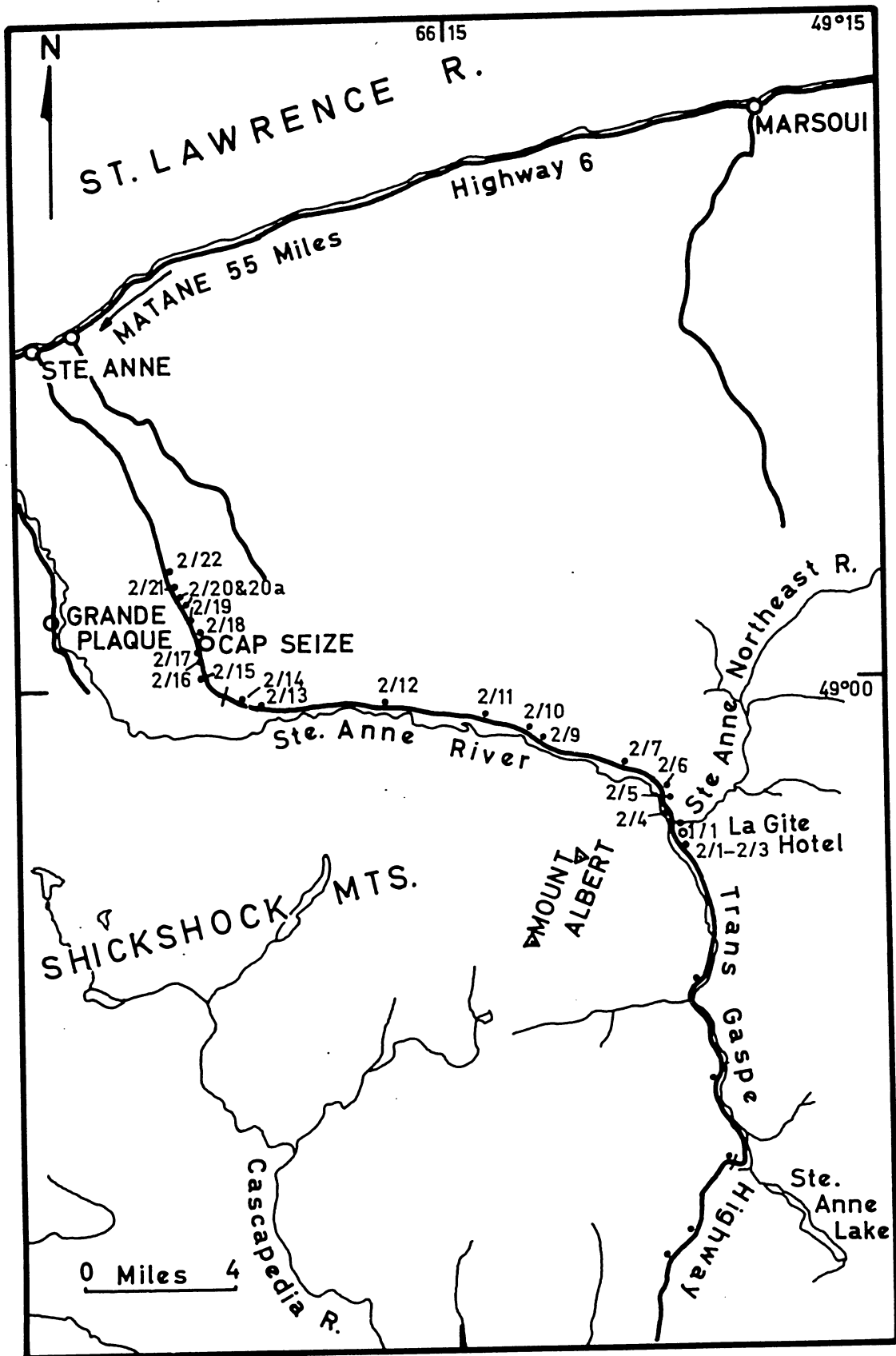
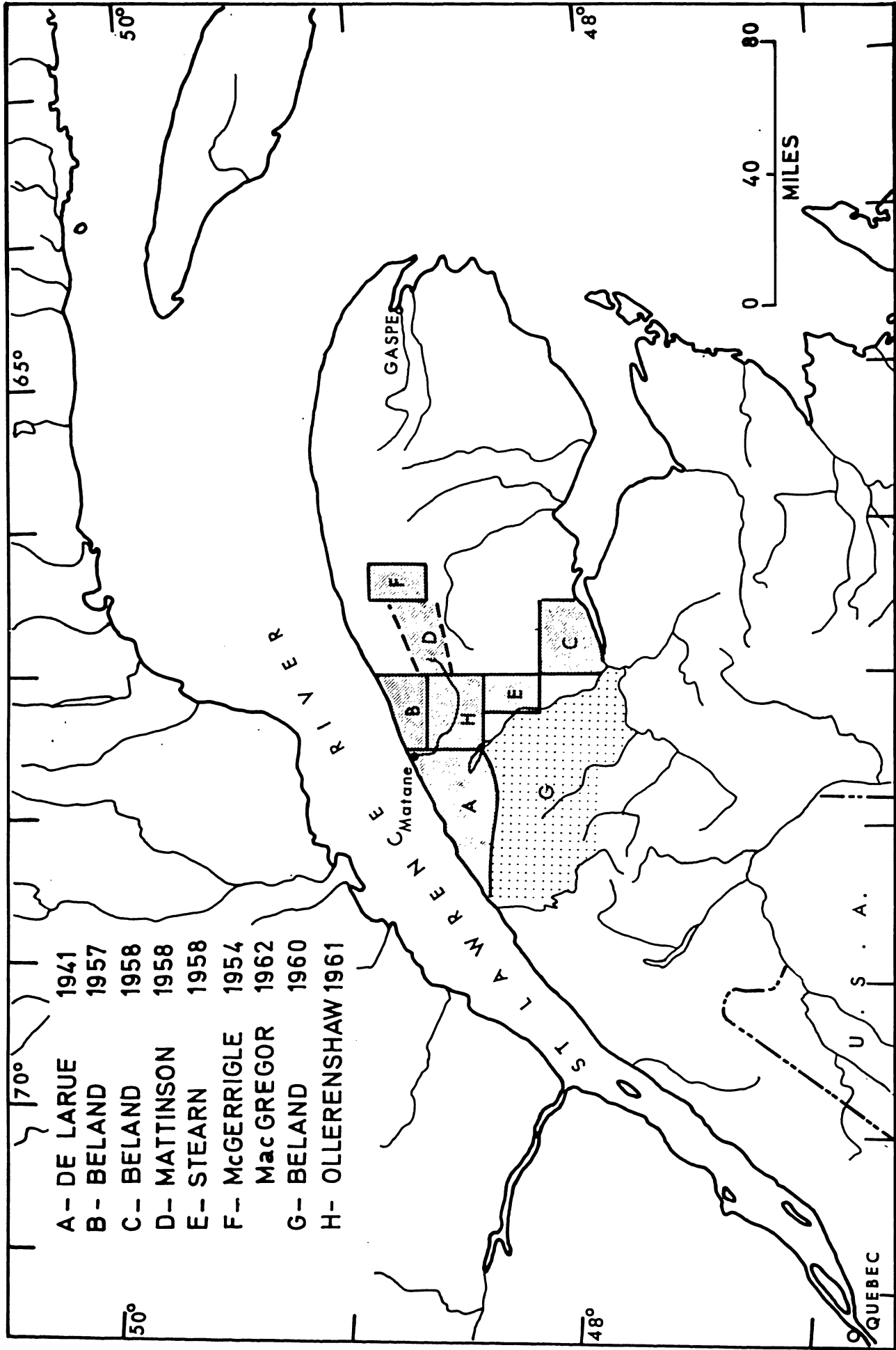


FIGURE 3. MAP SHOWING STRUCTURAL LOCALITIES IN STE. ANNE TRAVERSES NEAR MOUNT ALBERT



**FIGURE 4. MAP SHOWING AREAS INVESTIGATED BY PREVIOUS WORKERS**

Each outcrop of the rock types was studied in detail for structures of all scales. The readings were distributed randomly but carefully over the outcrop area. In the outcrops where structural features were exposed, any personal bias arising from the structural trends being normal or parallel to the line of sight were reduced by taking well distributed readings on sections normal to each other. The outcrops that provide the structural information were plotted on 1 inch to 1 mile maps or on aerial photographs, whichever was more convenient. In the cases where the outcrops were very small, the readings were collectively referred to a central locality.

On a regional scale the dominant structural trends make up approximately homogeneous domains. However, marked inhomogeneity in trend occurs within individual outcrops. It was thus impossible to divide the area into separate homogeneous domains with respect to each meso- or macro structure. The area, therefore, was arbitrarily divided up into 34 domains, each of 100 square miles and with little or no difference in structural trend from those adjacent.

In total, 28 weeks were spent in field in the summers of 1964-65. Three weeks were spent in the summer of 1966 to verify some previously made observations and to make fresh traverses along Mount Albert sections east of the study area (fig. 3). Most of the traverses were made alone by the worker. A field assistant was loaned to him by Dr. B. Skidmore of the Government of Quebec for seven weeks, the first half of the second field season.

## 2. History of Previous Work

The area was visited by Logan (1846, 1863) and Murray (1847) who traversed the Matane and Ste. Anne Rivers. Ellis (1885) and Low (1885) investigated the Matane River sections. Alcock (1926a) examined the rock exposures along the Matane and Petite Matane Rivers, and also in the Oak Bay area (1935), shown by area C in fig. 4.

Cambrian. Significant to the present work is an examination of <sup>He cc</sup> as associated with Matapedia Lake and Matapedia River sections by Crickmay (1932). He of the conglomerate carried out a regional synthesis of central western Gaspé, and en-possibility that visaged an Upper Ordovician-Lower Silurian unconformity between tightly succession of r

folded Cambro-Ordovician slates and volcanics and flat lying Silurian quartzite and limestone. He dated a belt of fine clastics and limestone as Upper Ordovician, the northern part which was later defined as Devonian (Beland, 1958). Hence the contact lying south of Causapsal village (fig. 6, in pocket) placed formerly between Silurian and Upper Ordovician, in fact separates the Silurian from Devonian rocks.

De laRue (1941, p. 15) investigated northwestern Gaspé, north of Matapedia Lake. He suspected an unconformity between the Sillery Formation of Lower Ordovician age, which consists of arkose, shale, sandstone, quartzite conglomerate and the Middle Ordovician Pohenegamook Formation of shale, slate, sandstone, and limestone. Another unconformable contact was proposed between the Middle Ordovician and Middle Silurian rocks.

McGerrigle (1954, p. 20) noted a parallelism between cleavage and bedding in the Shickshock rocks and suggested that a tectonic foliation (schistosity) was folded. A higher grade of metamorphism of the sedimentary and volcanic rocks to the south of those of Ordovician age indicated to McGerrigle that the northern boundary of the Shickshock Range is a high angle thrust. MacGregor (1962), Ollerenshaw (1963) and the present worker differ from this interpretation (see Structure of the Shickshock Mountain Area).

McGerrigle (1954, p. 23-24) suggested that a belt of complexly folded beds immediately north of the Shickshock Mountains was older than rocks farther to the north. "This belt is very complex structurally...reversed sections may be considered to be quite as common as normal ones in the belt". He attributed an age of 'Lower Ordovician or Older' to these rocks, basing his views on stratigraphic position and graptolitic fossils in Middle Ordovician rocks occurring to the north.

Within what he called 'Lower Ordovician or older', McGerrigle found limestone conglomerate containing Olenellids dated as Lower Cambrian. He considered these conglomerates, up to 50 feet in thickness, as associated with quartzites. Although he thought of the development of the conglomerate and quartzite as local, he did not rule out the possibility that the older sediments might be infolded with the confused succession of rocks.

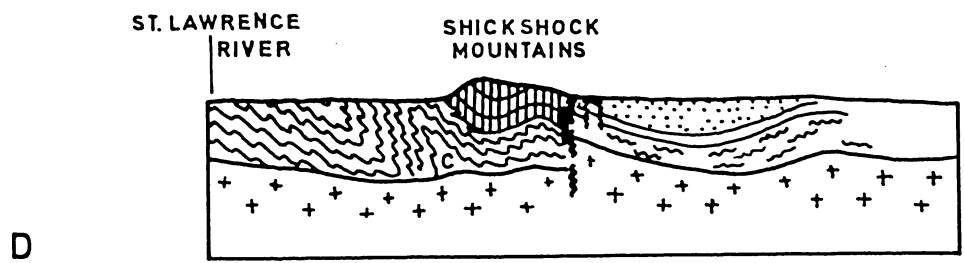
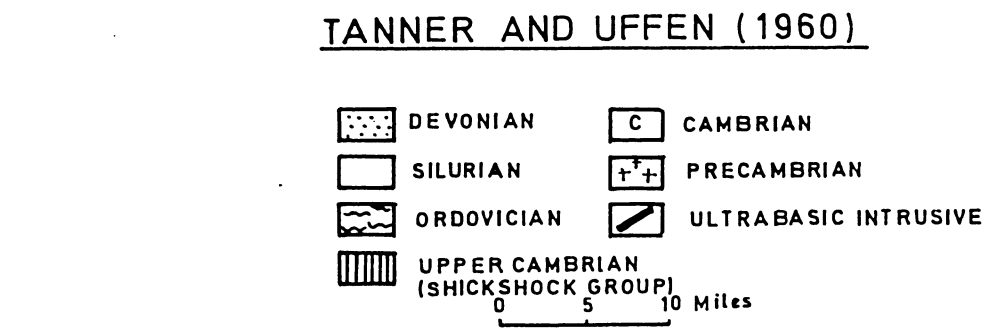
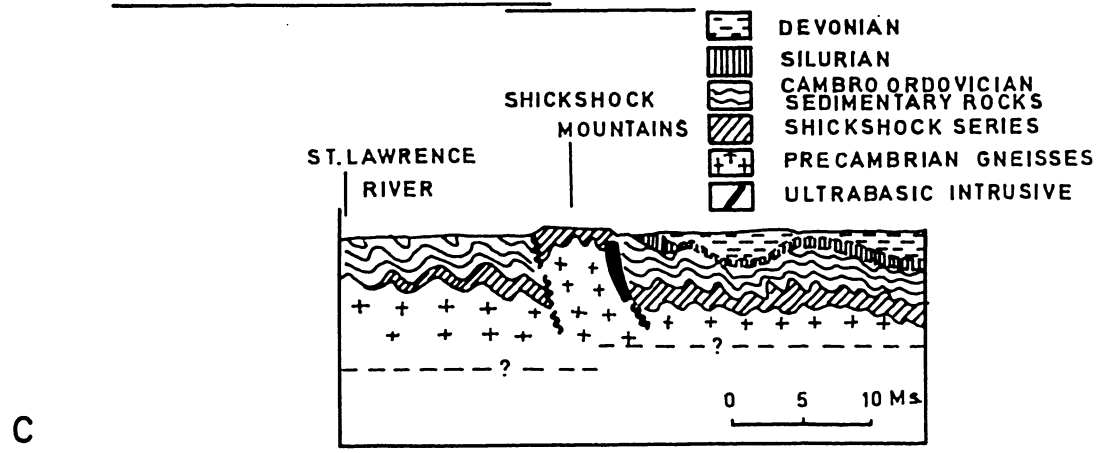
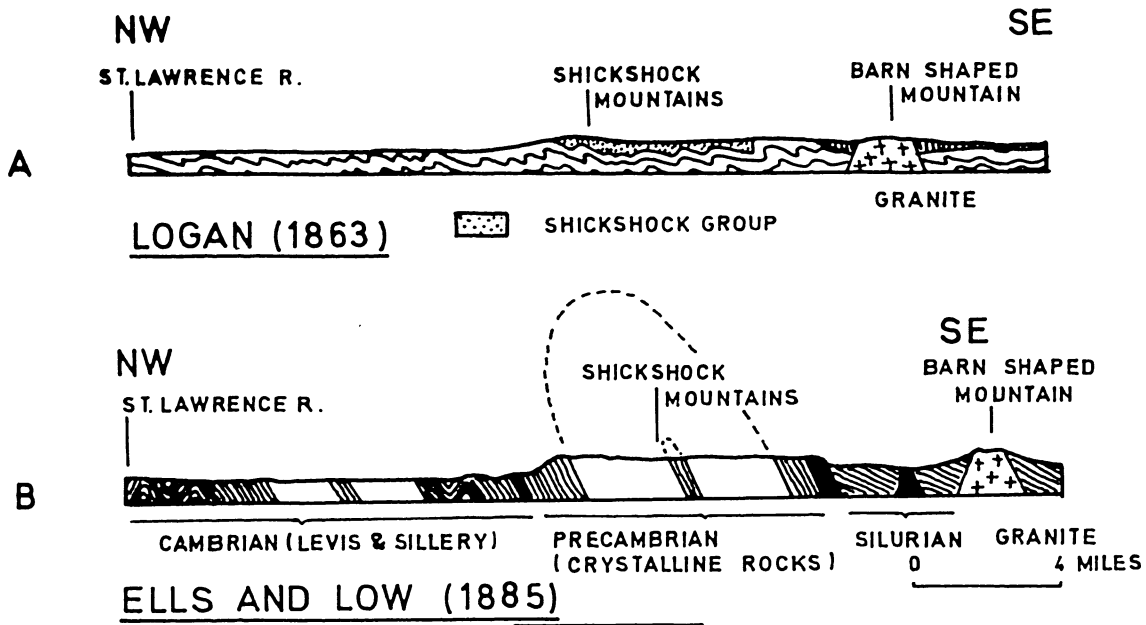
Beland (1957) mapped the Ste. Felicite-Grosses Roches area north of the Shickshock Mountains (area B, Fig. 4). A series of V or pod shaped patterns of quartzite outcrops that often face each other were mapped. He considered an impure grey sandstone, equivalent to the Pillar Sandstone of Tourelle and Courcellete areas situated 60 miles northeast of Matane (McGerrigle 1954), as the basal Ordovician bed north of the Shickshock Mountains. He also noted an increase of metamorphic grade southward (verbal communication 1964).

Beland (1958) mapped the Oak Bay area (fig. 4). He noted a common parallelism between bedding and cleavage in Upper Ordovician rocks suggesting tight to isoclinal folds. He distinguished the Fortin slate belt of Devonian age from the Matapedia Group of Upper Ordovician age. He also observed reef sediments and a volcanic rock association in the Siluro-Devonian rocks.

Mattinson (1958) carried out a petrological study of the Shickshock group of rocks in the Mount Logan area (area D, Fig. 4). North of the Shickshock Mountains he noted an increase in metamorphism towards the south in Cambrian-Ordovician slates and also a parallelism between bedding and schistosity. He suggested the presence of isoclinal folds in the rocks north of the Shickshock Mountains. Mattinson also observed a parallelism of schistosity and bedding in the matrix of sandstone exposed to the north of the Mountains. He noted the presence of a steep fracture cleavage which post-dates the schistosity. The occurrence of graptolitic Middle Ordovician slates faulted against the Shickshock rocks was also first reported by him (location shown in fig. 6).

MacGregor (1962) considered that the intrusion of the Mount Albert body to the northeast of the area postdates the schistosity in the Shickshock rocks. The contact aureole of the ultrabasic body was dated by the K-Ar method at  $495 \pm 35$  m.y. According to MacGregor a swing of the Schistosity within the Shickshock rocks around the igneous body is due to the intrusion. The radiogenic date matches with the date obtained by Ollerenshaw (see the following).

Ollerenshaw (1961, 1967) mapped the western Shickshock area in the Couq-Langis counties. He considered the Shickshock volcanic rocks as a wedge within the Quebec Group (Middle Ordovician).



**FIGURE 5. SCHEMATIC INTERPRETATION OF THE SHICKSHOCK MOUNTAINS BY DIFFERENT WORKERS SHOWING STRUCTURAL POSITION OF THE SHICKSHOCK GROUP**

According to him the Matapedia Lake volcanics are equivalent to the Shickshock metavolcanics, and the 'lower' contact with the slates is gradational. The biotite within the metasediments of the Shickshock Group from Duvivier Brook yielded a K-Ar age of  $530 \pm 35$  m.y. (location shown in fig. 6), which is Middle-Upper Cambrian, (Kulp 1961). Hence he suggested a pre-Taconic, possibly late Cambrian phase of deformation followed by a Devonian (Acadian) orogeny to account for later structures.

Beland (1960) in a study of the southwestern part of the area suggested that a Taconic (Late Ordovician-Early Silurian) unconformity separates the tightly folded slates of Cambro-Ordovician age from the gently folded Silurian rocks.

Beland (verbal communication 1964) indicated a locality near St. Cleophas southwest of Matapedia Lake where gently dipping Silurian sandstone is exposed within a quarter of a mile from an outcrop consisting of tightly folded slates of Cambrian-Ordovician (?) age. He also suggested the presence of basal Silurian conglomerate beds found only as gigantic boulders on the south shore of the Matapedia Lake. These were later called incognito conglomerates by Ollerenshaw (1963).

#### History of the interpretation of Structure of the Shickshock Mountains.

Logan (1863, p. 265) interpreted a synclinal structure for the Shickshock Mountain area with 'undulations along the middle'. According to his section (fig. 5A) the Shickshock rocks were younger than rocks exposed to the north of the Shickshock Mountains.

Ells (1885) and Low (1885) interpreted the structure as a tight overturned anticline of Precambrian rocks (fig. 5B).

McGerrigle (1954, p. 21-22) concluded that the structure of the Shickshock Mountains is more complex than previously supposed. He identified two anticlines and two or three synclines of foliation (schistosity) in the Mountains. He stated that in the Tourelle and Courcellette areas (location shown in fig. 4) "A test of validity of the assumption that bedding and foliation (schistosity) are parallel in general is provided by the band of sedimentaries bordering

Cascapedia Lake. This band has a northeast-southwest trend and schistosity trends likewise although dipping inward on either side of the belt... In southeast part of the map area the schistosity strikes N-S and dips to the east. These swings away from the general trend evidently are related to the Mount Albert serpentine mass whose western border lies to the east of the area. The fact that the schistosity here seems generally parallel to bedding may imply that it is strictly of bedding plane type, and it reacted to folding forces as, and with, the bedding. Or it may be that an original foliation established under one orogeny was folded by a later orogeny". McGerrigle was also impressed by the difference between the metamorphic grade of the Shickshock metavolcanics and the volcanics sporadically occurring in Ordovician rocks north of the Shickshock Mountains. The presence of Lower Ordovician or older rocks was also suggested to lie in a belt north of the Shickshock Mountains.

Mattinson (1958, 1964) considered that the Shickshock rocks were of the same age as the sedimentary rocks exposed to the north: "... The rocks for some distance north of the Shickshocks become generally younger than these. To explain the northward progression north of the Shickshock Group to the younger Normanskill and Levis rocks despite the generally southward dip, one must assume either an isoclinal overfold or a fault. In case of a overfold the Shickshock Group should reappear north of the fold axis. It does not.....". He avoided making a committing statement. The volcanic rocks, however, being non-persistent geographically do not have to re-appear if they were isoclinally folded.

Tanner and Uffen (1960) suggested from a gravity high over the Shickshock Mountain area that the Shickshock 'Series' were brought up as a horst (fig. 5C) of older rocks by two major faults on the northern and southern margins of the mountains. In their assumption the Shickshock 'Series' are the oldest rocks above the Precambrian gneisses.

The major objection against the interpretation is that a conformable northern contact separates the Shickshock Group from the grey slates, ruling out the possibility of faulted northern contact; if one assumes a normal sequence, the grey slates underlie the Shickshock Group.

Ollerenshaw (1963) interpreted the Shickshock Group as pre-Middle-Upper Cambrian in age. His deduction is contradictory with other evidence he presents as the Shickshock Group according to him is conformably overlying the Middle-Upper Cambrian grey slates (p. 193, fig. 14; shown in this thesis, fig. 5D). Although he indicates isoclinal folds that appear refolded in the Shickshock Group, he does not differentiate them from folds in the underlying rocks, exposed in a northerly belt, which in his section are shown as openly folded probably by a single phase of folding. His interpretation of a pre-Middle-Upper Cambrian age requires a faulted or thrust northern contact for the Shickshock Group. On the other hand, he himself advocates a conformable northern (lower) contact (1963, Geol. map, and p. 166-167).

### 3. Purpose of the Study

The study describes in detail the tectonic structures in the area and attempts to:

- i. Demonstrate that the earliest folds in the Cambrian-Ordovician rocks are pre-Middle Ordovician in age.
- ii. Demonstrate that the dominant folds in both post and pre-Silurian rocks are post-Middle Devonian in age.
- iii. Show that the Devonian (Acadian) fold elements are in all rock types that constitute the Gaspe Synclinorium and that the fanning of the corresponding cleavage is contemporaneous with folding.
- iv. Advance a new interpretation of the gross structure of the Shickshock Mountain area.
- v. Demonstrate the influence of lithology in folding.
- vi. Demonstrate the mechanism of folding.
- vii. Account for the curved axes of the Devonian folds.
- viii. Describe and present a classification of axial plane fracture cleavage.
- ix. Show that the joint planes in the folded rocks are independent of the attitudes of local Devonian (Acadian) fold plunges.

#### 4. Acknowledgements.

The worker is indebted to Dr. W. K. Fyson for his supervision, discussions and criticism. The work was carried out with the assistance of a grant from the National Advisory Committee for research in Geological Sciences, and it is partly sponsored by the Department of Geology, University of Ottawa. It is also a part of 'Structural relations in the Maritimes' a project carried out under Dr. W. K. Fyson. The help provided by the Geology Department, Ministry of Natural Resources of the Government of Quebec is gratefully acknowledged. The worker wishes to thank Dr. B. Skidmore and Dr. C. Hubert for discussions and assistance. Acknowledgement is also due to Professor J. Beland of the University of Montreal for constructive discussions. The worker is grateful to all the forest wardens who extended every assistance during the field work; and Mr. F. Bernard, field assistant, on loan from Dr. Skidmore.

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## II GEOLOGICAL BACKGROUND

1. Stratigraphy

For individual rock description see Appendix 1, Page 97

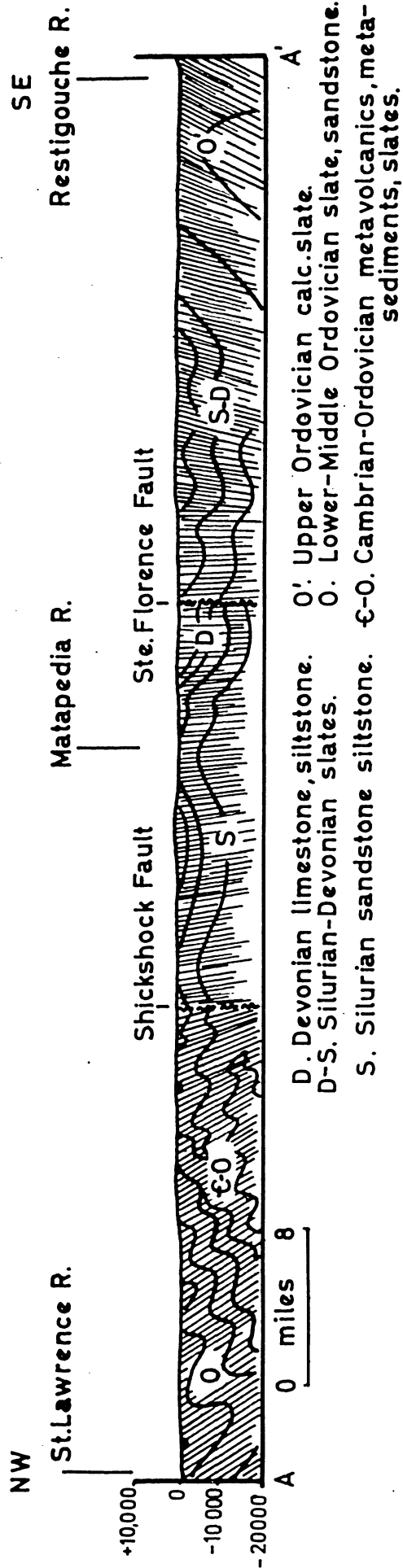
Table 1. TABLE OF FORMATIONS

ERA SYSTEM	PERIOD	FORMATION OR GROUP (Thickness in Feet)	DESCRIPTION
CENOZOIC	Pleistocene and recent		Boulder Clay, clays, silts, sands gravels.
PALEOZOIC Devonian	Middle and Lower	York River-Heppel Formations 12,000	Grey calcareous sandstone, red sandstone, calcareous siltstone, minor conglom- erate.
	Middle and/ or Lower	Fortin Group ? Ste. Marguerite volcanic member. 700	Slates, Calcareous and phyllitic, thin layers of siltstone, sandstone. Beds of greywacke. Amygdaloidal and compact andesitic volcanics, minor breccia.
	Lower	Grande Greve For- mation 2,500	Calcareous siltstone, silty limestone.
		Cape Bon Ami For- mation 2,500	Argillaceous and silty limestone occasionally calcareous shale.
	Silurian <sup>1.</sup>	Upper	St. Leon Formation 5,000
Middle		Syabec Formation 500	Wavy bedded fossiliferous shaly, silty limestone.

ERA SYSTEM	PERIOD	FORMATION OR GROUP (Thickness in feet)	DESCRIPTION
Silurian	Middle	Val Brillant Formation 500	Orthoquartzitic sandstone calcareous cement in upper part.
		Awantjish Formation 150- 1000	Fossiliferous shale, siltstone, sandstone, conglomerate
Ordovician	Upper	Matapedia Group <sup>2</sup> .	Grey to dark grey, dense calcareous slate.
	Middle	Quebec Group	Slates, grey, dark grey, red green, calcareous, silty, quartzite conglomerate (Kamouraska facies), limestone and impure grey sandstone.
	Lower		
Cambrian	Upper	Cambrian-Ordovician rocks	Grey, dark grey, red, green calcareous silty, phyllitic, sericitic slate with limestone, sandstone, quartzite, sheared conglomerate.
		Shickshock Group 14,000 <sup>3</sup> .	Metamorphosed arkosic sandstone, metaarkose, minor phyllitic slate, Basalts, metavolcanics.

Thick vertical line indicates precise paleontological dating of rocks, dashed line indicates paleontologically deduced approximate age; and absence of the line suggests absence of paleontological age evidence.

1. 'Silurian and Devonian' in the southern part of the area includes the Chaleur Bay Series made up of calcareous shale, siltstone, sandstone, silty sandy or shaly limestone, coralline and reef limestone, conglomerate; basic, intermediate and acidic volcanics, tuff and agglomerate, intruded by acidic and basic dykes and sills. The Upper Ordovician, Silurian, and Devonian rocks make up a conformable rock succession in the southern part of the area.
2. The Upper Ordovician is missing from the northern part of the area. The Matapedia Group was regarded as containing presently known Fortin Group (Alcock 1926). The Fortin Group is considered as facies equivalent of more northerly occurring Lower and/or Middle Devonian rocks.
3. Thickness from Ollerenshaw (1961). This deduction is doubtful since Ollerenshaw estimated the thickness from the attitudes of schistosity rather than those of bedding. See Appendix I (p. 102) for the age of the Shickshock Group. See Age of the Cambrian-Ordovician Slates (Appendix I, p. 104) for the Middle-Late Cambrian trilobite locality (also fig. 6). See Age of  $F_1$  Folds (Chapter III, p. 31) for the age of Tamagodi Sandstone (location shown on fig. 6). See Appendix V for the age of the Middle Ordovician slate outcrop in the Shickshock fault zone, (location shown on fig. 6). The ages of Ordovician, Silurian and Devonian sedimentary rocks are based on the conclusions of previous workers on fossil evidence (Appendix I).



**FIGURE 6b** S<sub>2</sub> AXIAL PLANE CLEAVAGE FAN SHOWN APPROXIMATELY ALONG SECTION A-A'  
 (GEOLOGICAL MAP FIG. 6) ACROSS GASPE SYNCLINORIUM

## 2. Taconic Problem

In many areas in the Appalachians an unconformable boundary is placed between tightly folded Middle-Upper Ordovician slates and limestones and openly folded Silurian quartzite and limestone.

This structural disharmony and low angled thrusts and nappes of immense magnitude account for abnormal distributions of the Ordovician sedimentary facies, and are considered as manifestations of the Taconic Orogeny.

Logan (1843, 1863) first invoked post-Ordovician and pre-Silurian deformation in the Canadian Appalachians. He proposed an extensive pre-Silurian folding and thrusting of Ordovician slate, quartzite, and limestone sequences. He traced a belt of folds, characterised by low angled thrusts on the western and northwestern margin from Newfoundland to southern Quebec. Consequently, the line in Quebec that separates folded Ordovician clastic rocks from flat lying Ordovician limestone and sandstone to the west is called Logan's line. Further ideas on the Taconic Orogeny were subsequently developed by geologists in New York and Vermont.

In this discussion the Taconic Problem refers mainly to the stratigraphic position of the Taconic Orogeny in the Ordovician system. However the Taconic Problem also refers to the presence or absence of klippen to account for allegedly anomalous distributions of Ordovician shale in a limestone-quartzite facies. Therefore some account of the klippen hypothesis is necessary.

### Klippen Theory

Linear carbonate outcrops enclose shale facies from Champlain Lake to Connecticut on the east, and from northwest of Rutland, Vermont to as far south as Saratoga, New York on the West (Rodgers 1937, Dale 1898, 1904, 1923). Ruedmann (1909) interpreted large scale thrusts between the Snake Hill and Canajoharie shale formations overlying the Bald Eagle carbonate sequence in the Taconic region. This was not totally agreed upon by later workers (Prindle and Knopf 1932, Kay 1937, Cady 1945) who suggested that the Snake Hill and Canajoharie are correlative and are not thrust on each other. According to Kay (1937), the limestone conglomerate used as important evidence

Table 2. Position of Unconformities in Cambrian and Ordovician  
In the Northern Appalachians (Modified after Zen, 1961)

		N.W. Taconics	N. Central Vermont	S. Quebec	Quebec City	N. Central Gaspé	Western Gaspé	Suggested for Western Gaspé
		Cady 1945 Zen 1961	Cady 1956 Albee 1957	Cooke 1950 Osberg 1956	Rasetti 1946 Osborne 1956	McGerrigle 1954	MacGregor 1962 Ollerenshaw 1963	
Ordovician	Middle	Hortonville and Ira Pawlet ~~~~~	Mortown	Beauceville ~ ~ ~	Quebec City	"Middle Ordovician"	Matane River Group	Quebec Group ~~~~~
	Lower	Mount Hamilton		Caldwell 	Levis	Pillar Sandstone		
Cambrian	Upper		~~~~~ Stowe Ottaugeuchee	Conformable 	Lauzon	"Lower Ordovician and older"	~~~~~	
	Middle	~~~~~ West Castleton	Conformable					
	Lower	Bull Biddie Knob	Camels Hump	Sutton Schists	~~~~~ Charny	Contact faulted   Shickshock Series	Matane River Group (with Shickshock Group)	Cambrian-Ordovician rocks (with Shickshock Group)

could not be a fault breccia because the shales could not shed carbonate conglomerate during transport. There have been recent doubts about the presence of extensive nappe structures in the American Appalachians. Platt (1962) considered the western part of the "klippe" to be autochthonous. Craddock (1957) rejected the presence of nappes in the Kinderhook Quadrangle N.Y. on grounds of insufficient evidence. He noted tight folds in competent rocks and open folds in the incompetent ones to account for apparent structural contrasts. Zen (1961) in an area in west central Vermont, assumes the presence of folded nappe structures. The thrusting is invoked to have resulted from a submarine gravity mechanism.

In the Appalachians of southern Quebec, evidence to substantiate the Klippen theory is fragmentary, but contrasting fold styles in rocks may indicate the presence of nappes in the Sherbrooke area between Ordovician and older rocks (St. Julien, verbal communication 1967). Rodgers and Neale (1963) and Tuke (1966) consider nappe structures in Newfoundland to account for the presence of Lower Ordovician and older shales and greywackes structurally overlying Middle Ordovician carbonates.

#### Stratigraphic Position of Ordovician-Silurian Unconformities and Orogenies

Schuchert (1930) maintained that intense folding and overthrusting of rocks near Lake Champlain in Vermont along the northern Appalachians are late Ordovician in age (as also outlined by Logan 1863, Dana 1888, Alcock 1926a). He also concluded that whereas in Vermont, southern Quebec and New York folded Taconic rocks were not refolded but only epierogenically elevated and normally faulted by Devonian crustal deformations, the southeastern edge of the geosyncline was refolded and intruded by granites.

Other evidence for an Upper Ordovician orogeny according to Schuchert (1930, p. 712) is the absence of ultrabasic bodies in Siluro-Devonian rocks near Mount Albert in Gaspé (Alcock, 1926b), where the ultrabasic rocks intrude the pre-Silurian Shickshock volcanic rocks.

Schuchert (1930) also observed that in the Matapedia Valley, low dipping Middle Silurian Val Brillant quartzites overlie

a tightly folded succession of slate and volcanic rocks that were assigned to the Ordovician. There is evidence now that the folded rocks are Upper Cambrian or older (Ollerenshaw 1963, MacGregor 1962). Hence the unconformity, if it exists, could represent a deformation at any time during the Ordovician. A faulted contact may also be suggested.

Crickmay (1932) agreed with Schuchert on the presence of a 'Taconic' angular unconformity in the Matapedia Lake area where he compared the structures in the Silurian and Ordovician rocks. Near Causapscaal village he invoked an unconformity that represented the Taconic Orogeny and separated a tightly folded supposedly Cambro-Ordovician slate from Siluro-Devonian massive limestone and sandstone. He noted (p. 382) "...the complete lack of any cleavage in Silurian rocks and their slightly metamorphosed character mark them as having suffered little as a result of orogenic movements and certainly preclude any idea that they form a thrust plate. There is no escape from the conclusion that the complex structures of the Quebec Group (slate-limestone-quartzite-limestone breccia sequence of Cambro-Ordovician age) are pre-Silurian and originated during the Taconic orogenic epoch....". Thus Crickmay suggested that a Late Ordovician orogeny took place in Matapedia Valley area.

Beland (1958) revised the Matapedia section and found that the tightly folded rocks treated as Ordovician by Crickmay are in fact Fortin slates of Early Middle Devonian age.

So far as the observations on the style of folds in the two lithologies are concerned, Crickmay was correct. Therefore, within the conformable sequence of massive Siluro-Devonian quartzites and the laminated slate severe disharmony in fold style occurs; whereas the folds are tight to isoclinal in slate, they are open in the massive limestone and quartzite. It is hence suggested that the competency of the rocks deformed is a significant factor in the abrupt change in fold style.

A pre-Middle Ordovician unconformity is suggested for many areas by recent work (Table 2). Riordon (1957), in the Thetford-Disraeli area of southern Quebec, indicated an angular unconformity between Ordovician and Silurian rocks, also assumed by Cooke (1950).

He strongly suspected that the Sutton-Bennet schists (probably equivalent in age to the Shickshock meta-volcanic schists in the Gaspé area) and the overlying Caldwell Group, a complexly folded Cambro-Ordovician sequence of slate, limestone and quartzite were conformable. These were deformed by an early Ordovician orogeny before being unconformably overlain by the Beauceville Group of Middle Ordovician age.

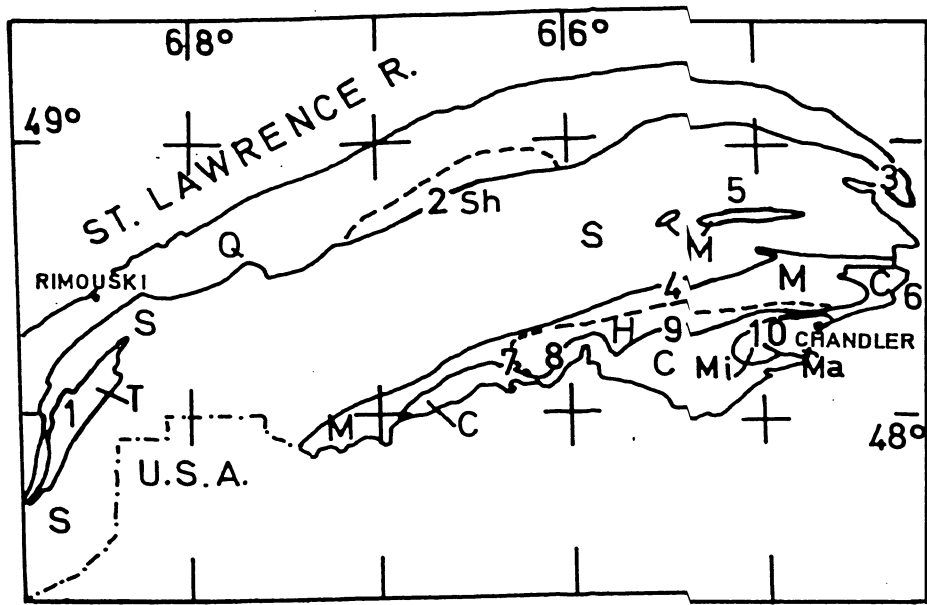
Structural analysis in these complexly folded rocks is extremely difficult, consequently the conclusions of the workers are often contradictory. Baer (1963) in the Memphramagog area of southern Quebec regarded all structures in Lower Paleozoic rocks as younger than the Devonian rocks in the area.

According to him the main phase of folding was Acadian in age and an Upper Ordovician-Lower Silurian deformation was limited to slight warping.

In the nearby Sherbrooke area (St. Julien and Lamarche 1965) an angular unconformity separates Lower Ordovician schist, siltstone, phyllite, sandstone and acidic volcanics from Middle Ordovician slate, sandstone and limestone. On the unpublished map of the Sherbrooke area (St. Julien, verbal communication, 1967) the Cambrian or Lower Ordovician rocks occasionally strike N-S, whereas the adjacent overlying Upper Ordovician rocks strike NE. The Middle Ordovician is unconformably overlain by Silurian sandstone, limestone and conglomerate. Both unconformities were folded by a Devonian (Acadian) deformation.

Burke (1964) indicated a sub-Silurian unconformity at five localities in eastern and southern Gaspé peninsula. In only one locality, near St. John River in eastern Gaspé, do Silurian rocks appear to directly overlie Upper Ordovician rocks; their relationship is not very clear but "appears unconformable" (p. 456). In other places the Silurian lies directly on folded Lower Ordovician to Upper Cambrian (Burke 1964, p. 455; McGerrigle 1950, p. 27), so that as in the Matapédia Valley, deformation could have been at anytime during the Ordovician.

McGerrigle (1954) considered the Shickshock meta-volcanic schists to be Cambrian or older. He invoked (p. 22) folding of an



- C- Silurian and younger rocks.
- S- Central Synclinal Siluro-Devonian
- M- Matapedia Group
- H- Honorat Group
- Mi- Mictaw Group
- T- Trinite Group
- Q- Quebec Group
- Sh- Shickshock Group
- Ma- Maquereau Group

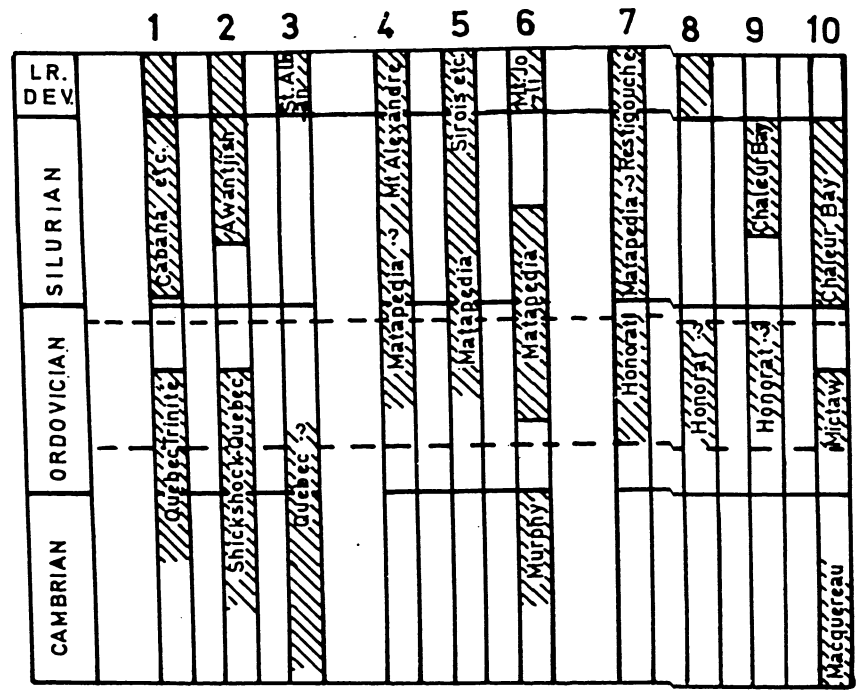


FIGURE 7. STRATIGRAPHIC COLUMNS FOR LOCALITIES SHOWN IN MAP OF GASPE REGION. AFTER SKIDMORE (1967, MS.)

early schistosity in the Shickshock rocks by late folds. Immediately to the north of the Shickshock Mountains he observed a complexly folded belt of Lower Ordovician or older slate, limestone and quartzite. He inferred that these rocks had been deformed before the Middle Ordovician and were represented further north by less deformed fossiliferous rocks.

Mattinson (1958) traced the folded schistosity in the Shickshock meta-volcanic rocks into the slate and sandstone immediately to the north of the Mountains. He suggested a conformable northern contact between the Shickshock metavolcanics and the slates. This folded schistosity is absent from the Middle Ordovician slates of the Quebec Group further north. Such a conformable northern contact was also supported by MacGregor (1962) near Mount Albert and Ollerenshaw (1963) along the Matane River and the Matapedia Lake.

Although Mattinson (1958) does not forward any tentative conclusions as to the structural disharmonies in the area, it is obvious from his observations that there is a marked difference in the fold style in the Lower Ordovician-Upper Cambrian and the Middle Ordovician rocks.

Skidmore (abstract, 1967, and manuscript, The Taconic Unconformity in Gaspé Peninsula and Neighbouring Regions, 1967) reviews the Ordovician deformations in Gaspé Peninsula. His views are summarised in figure 7.

For western Gaspé in the northern part of the area (locality 2, fig. 7) the stratigraphic column shows a gap representing the Late Ordovician and Early Silurian. In the southern part of the area (west of locality 7, fig. 7a) stratigraphic gap probably occurs in pre-Early Ordovician times, but the succession is continuous from Upper Ordovician to Devonian. This indicates that, whereas the northern part of the area lying towards the margin of the sedimentary basin was tectonically active in the Lower Paleozoic, the southern part situated farther away was a site of continued deposition.

A pre-Ordovician unconformity is present in southeastern Gaspé near Chandler (fig. 7), but its extent is not yet known.

In the thesis area in western Gaspé, the only indications of an Upper Ordovician unconformity are the conglomerate beds seen

as boulders, but never found in situ, on the shores of Matapedia Lake, near Val Brilliant, called "incognito conglomerates" by Ollerenshaw (1963) and dated as Middle Silurian. Beland (verbal communication 1964) referred to a locality near St. Cleophas, 8 miles southwest of Lake Matapedia, where within half a mile separates tightly folded Ordovician slate and Silurian sandstone dipping at  $45^{\circ}$ . A probable unconformable contact is suggested, but the age of the slate is open to question.

#### Radiogenic Dating of the Ordovician Deformations

Radiogenic dating of the unfossiliferous rocks has contributed immensely to unravelling of the ages of the deformed rocks.

Muscovite considered metamorphic in the schistose sandstone (with a schistosity in the matrix parallel to bedding) in the Shickshock Group gives a minimum K-Ar date of deformation at  $530 \pm 35$  m.y. (Ollerenshaw, 1963; Lowdon et al., 1962). (See fig. 6 for location) According to Kulp's scale (1961) this is Upper-Middle Cambrian. Since Ollerenshaw regards the Shickshock Group of rocks to be interbedded with grey slates dated by him as Cambrian, he concludes that the main deformation took place in the Upper-Middle Cambrian.

The contact aureole of the Mount Albert ultrabasic intrusion east of the area yielded a minimum K-Ar age of  $495 \pm 35$  m.y. which is Early Ordovician on the time scale (MacGregor, 1962; Lowdon et al., 1962). It is considered that the intrusion was later than the development of the schistosity in the Shickshock rocks. This is suggested by the deflection of schistosity around the igneous body, and lack of cataclasis within it. Layering in the igneous body trends transverse to the schistosity in the host rock (MacGregor, 1962; Smith, C.H., verbal communication 1965, see Appendix III).

Granites that inject into an ultrabasic body which in turn appears to intrude the Beauceville Group of Middle Ordovician in wouthern Quebec gave a K-Ar date of 477-481 m.y., close to the boundary of Early Ordovician on Kulp's scale, (Leech et al., 1963) (see below for comments).

However, another group of radiogenic dates indicate an age of crustal deformation younger than the Middle Ordovician. A supposed authigenic biotite in the Upper Ordovician Utica Shale, north of

Montreal gave an age of  $402 \pm 25$  m.y. (Beall 1962). Biotite in the Jacques Cartier granite in Gaspé crossing the Lower Ordovician sediments yielded a minimum age of  $420 \pm 30$  m.y. (Leech et al., 1963). These can be interpreted as latest Ordovician-Early Silurian dates according to Kulp's scale.

Rickard (1965) considered that dating of mica recrystallised along a late stage crenulation cleavage that deforms an existing schistosity would date the last pulse of deformation. In Cambrian rocks and Sutton schists near Sutton, southern Quebec, these micas yielded an age of 420-440 m.y. Accordingly, the last deformation was probably Late Ordovician in age.

Caldwell schists of Cambrian-Ordovician age, exposed 80 miles north of the Sutton area, are dated as  $414 \pm 30$  m.y. (Early Silurian) by a similar K-Ar method on crenulation cleavage (Leech et al., 1963).

The above group of radiogenic dates are considered to represent a deformation which reset the radiogenic clock in the older rocks, probably already deformed, back to zero, by degassing argon from the micas. The dates lying between 410-440 m.y., are considered to indicate a Late Ordovician-Early Silurian deformation.

A third set of radiogenic age determinations on the granites and Lower Paleozoic rocks along the Appalachians gives a date of 350-390 m.y. This represents the Acadian phase of deformation and igneous intrusion, which is considered post-Middle Devonian in the eastern Appalachians (Leech et al., 1963).

The two sets of radiogenic dates of 480-530 m.y. (Early Ordovician-Cambrian on Kulp's scale) and 410-440 m.y. (Early Silurian-Late Ordovician) have both been referred to as indicating a Taconic Orogeny. There are also controversial age relationships between dates of granite emplacements. Poole et al., (1963) proposed that since the granites that intrude the ultrabasics in southern Quebec date around 480 m.y. (Upper boundary of Lower Ordovician on Kulp's scale), they must be younger than the Beauceville Group, into which the ultrabasics are intruded. The Beauceville Group is known paleontologically to be Middle Ordovician (Cooke, 1950). Therefore, the limits of the Middle-Upper Ordovician should be lowered from 445

m.y. in Kulp's scale. Rickard (1964b), however, maintained that the ultrabasics intrude along an unconformity at the base of Beauceville Formation (used instead of group by Rickard), therefore the dates could indicate the minimum age of the lower part of the formation. He maintained that only the upper part of the Beauceville is dated paleontologically as Middle Ordovician; the lower part could be as old as the Early Ordovician (480 m.y.) and accommodate the dates of Poole et al., (1963).

Poole et al., (1964), appeared convinced that their dated granite is younger than the Beauceville, but the relationships are still obscure (Poole, verbal communication 1967).

### Summary

The main ideas regarding periods of deformation in the northern Appalachians prior to the Mid-Devonian Acadian Orogeny are as follows:

1. That a major deformation was pre-Middle Ordovician (Riordon 1957, McGerrigle 1954, MacGregor 1962, Ollerenshaw 1963, Smith, C.H. personal communication 1965, St. Julien and Lamarche 1965).
2. That a major phase of deformation of the rocks was Upper Ordovician-Lower Silurian (Logan 1863, Dana 1888, Ruedmann 1909, Alcock 1926a, Schuchert 1930, Crickmay 1932, Cady 1945, Craddock 1957, Beland 1960, Osberg 1965, Rickard 1965).
3. That a phase of deformation dated as 480 m.y. and at the Lower-Middle Ordovician boundary on Kulp's time scale is in fact Upper Ordovician, and hence the boundary of the Upper-Middle Ordovician be lowered to the present Lower-Middle Ordovician (Poole et al., 1964).
4. It has also been suggested that all the structures in Lower Paleozoic rocks are Devonian (Acadian) in age (Baer, 1963).

Another possibility that explains the varying ages of deformation in various areas is the formation of successive orogenic fronts which probably escaped deformation during later orogenic phases. Rickard (personal communication 1966) regards the areas in which older radiogenic dates are obtained because they escaped later deformation, as "tectonic peninsulas". In these areas, the structures as well as

radiogenic dates indicate an age different to that of the rocks of corresponding age elsewhere. The presence of "tectonic peninsulas" or "islands" could also mean that whereas in one region an Ordovician deformation is extensive and the Devonian orogeny minor in magnitude, in another, the magnitudes may be completely reversed.

In the northern part of western Gaspé, an Upper Ordovician-Lower Silurian phase of disturbance is suggested by the absence of Upper Ordovician and most of the Lower Silurian rocks, signifying uplift and erosion. It is marked in places, by a basal conglomerate bed (Ollerenshaw 1963). Near Trois Pistoles, 90 miles southwest of Matane, some disturbance is probably indicated by an angular discordance (Beland, verbal communication 1966) between Ordovician slate and Silurian sandstone. However, these unconformable relationships do not necessarily indicate a severe phase of folding in Late Ordovician or Early Silurian times.

An older more intense phase of deformation and metamorphism during or before the Lower Ordovician is correlative with deformation and metamorphism in other parts of the Appalachians.

### 3. Major Structural Features

#### Gaspé Synclinorium

The rocks in the area are folded into a synclinorium called the Gaspé Synclinorium in which Cambrian and Ordovician rocks flank the northwestern margin, and Upper Ordovician and Siluro-Devonian rocks are exposed on the southeastern part. Devonian rocks are exposed in the central part of the area. The formations in this structure are exposed as bands trending northeast (fig. 6).

The Shickshock Fault is a major structural element that varies in trend between NE and ENE and divides the area into distinct major divisions. The first lies between the fault and the St. Lawrence River, an 18-24 mile wide belt in which Ordovician and Cambrian-Ordovician rocks are exposed. To the south of the fault Siluro-Devonian and Upper Ordovician rocks outcrop. The undivided Cambrian

Ordovician rocks consist of phyllitic slate, sandstone, limestone with volcanics, and the Shickshock Group consisting of metavolcanic and metasedimentary rocks, lying in a band 8-9 miles wide.

Contact between the Shickshock Group and the Cambrian-Ordovician rocks to the north is considered conformable owing to (i) an interbedded contact between the Shickshock metavolcanic rock and grey slate (Appendix 111, Loc. St. A. 2/1-2/3), (ii) lack of shearing or tectonic disturbance along the contact, (iii) presence of slate bands within volcanic rocks of the Shickshock Group. Lying to the north of the Cambrian-Ordovician slates are Early and Middle Ordovician rocks of the Quebec Group. These consist of grey calcareous slate with siltstone, sandstone and limestone interbeds of variable thickness. These also contain quartzite and massive quartzite-conglomerates, which outcrop in discontinuous linear patterns.

A series of major folds in Siluro-Devonian rocks crop out between the Shickshock Fault and Ste. Florence Fault 22 miles to the south (fig. 6 and 6a). Fine clastic rocks of Lower and/or Middle Devonian age, the Fortin Group of slates, occupy the southern part of the axial area of the synclinorium. These are considered facies equivalents of the Gaspé limestone and sandstone groups of Lower or Middle Devonian age exposed 60 miles northeast of the area (McGerrigle 1954).

The Upper Ordovician, Silurian and Devonian rocks are conformable in the southern part of the area. The Fortin Group of Early and/or Middle Devonian age thus lies adjacent to the Matapedia Group of Late Ordovician age, separated by a band of unnamed Siluro-Devonian siltstone and sandstone. This is succeeded to the south by Siluro-Devonian clastic and volcanic rocks.

The Cambrian-Ordovician rocks in the north are presumably folded into an anticline. Ollerenshaw (1963) reaches the same interpretation in his structural sections in which he considered a central belt of older rocks exposed 'somewhere' in the middle of the Cambrian Ordovician belt of rocks (fig. 5).

The synclinorium involves all rocks older than the upper part of the Middle Devonian and it is considered as a Middle Devonian (Acadian) structure. A cleavage coincides with the axial planes of the small scale folds and there is a fanning of the axial planes

and cleavage from a southeasterly dip in the northern part of the area to a northwest dip in the southern part of the area (fig. 6b).

### Unconformities

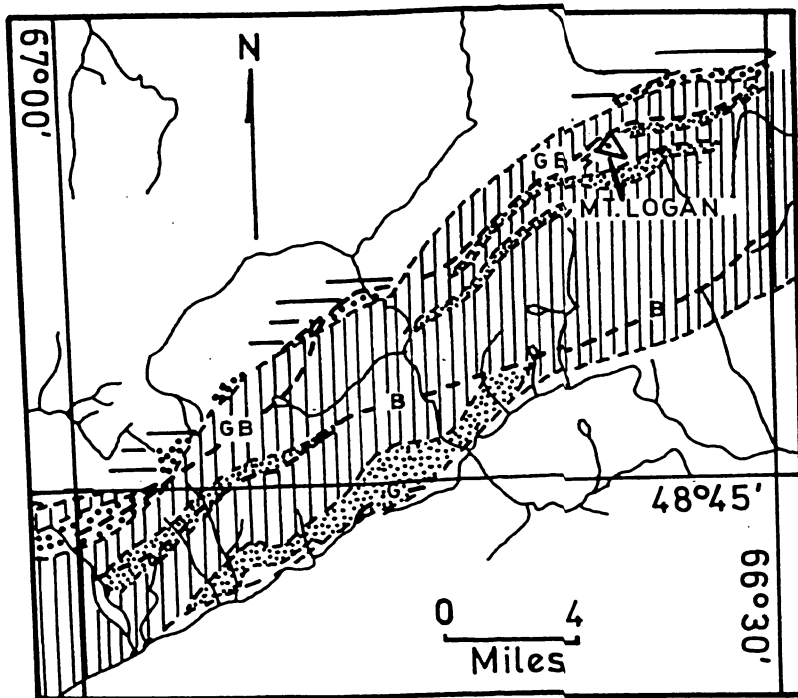
Three widespread unconformities are considered to be present in the Lower Paleozoic succession in the area:

- I. Between Cambrian-Ordovician rocks and the Lower Ordovician rocks. The phase of deformation comprised of intense folding ( $F_1$ ) of the Cambrian and some Lower Ordovician rocks (see p. 29).
- II. Between the Middle Ordovician and Middle Silurian rocks in the northern part of the area. The deformation that gave rise to the unconformity consisted of uplift and erosion and probably gentle warping accompanied with fault movements (see Taconic Problem, p.12 and Shickshock Fault p.71 ). In the southern part of the area the succession between Ordovician, Silurian and Devonian rocks is conformable.
- III. Between the Middle Devonian and later sediments which consist mostly of Recent gravel and sand. The deformation of the pre-Middle Devonian rocks formed tight to open folds ( $F_2$ ) throughout the area.

### Regional Metamorphism

Beland (1957) noted a southward increase in metamorphic grade of the Cambrian-Ordovician slates in Grosses Roches area north of the Shickshock Mountains (area B. fig. 4). A similar higher grade is observed in the southern parts of the slate belt of Cambrian-Ordovician age along the Matane and Cap Chat Rivers (fig. 6) and Ste. Anne River (fig. 3). South and southeast of Portage Lake (fig. 6) a higher metamorphic grade than to the north is apparent in the slates. These rocks are phyllitic, often highly sericitic and have a silvery lustre due to recrystallised micas on the schistosity surfaces. Calcareous or silty argillites are welded and not readily broken. The recrystallised appearance is not common in the northern parts of the slate belt, where the rock is dark to grey, with a dull lustre on cleavage surfaces and argillites resemble fractured mudstone.

Field observations indicate that the line which separates rocks with a higher metamorphic grade from those of lower grade can



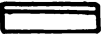

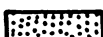

- |   |                       |                    |
|---|-----------------------|--------------------|
|  | Slaty Shale           |                    |
|  | Meta arkoses          | } Shickshock Group |
|  | Metasediments         |                    |
|  | Metavolcanics         |                    |
| --GB--  | Green biotite isograd |                    |
| --B--   | Biotite isograd       |                    |
| -G-   | Garnet isograd        |                    |

FIGURE 8. METAMORPHIC ISOGRADS IN MOUNT LOGAN AREA.  
AFTER MATTINSON (1964).

be drawn between Matapedia and Portage lake in the central part of the area, north of Riviere Matane village along Matane River, 4 miles north of the Shickshock Mountains along Cap Chat River; and 55 miles northeast of Matane along the Ste. Anne River, north of Cap Seize and south of Grande Plaque villages (fig. 3). The change in metamorphic grade is accompanied by development of a schistosity which parallels the axial planes of near recumbent isoclinal folds (photo. 1).

Mattinson (1964) drew isograds on a map of the Shickshock rocks near Mount Logan, 8 miles east of the study area (fig. 8). He showed an increase from the green biotite facies in the north, to garnet facies in the south, where it is cut off by the Shickshock Fault. The isograds are slightly oblique to the predominant outcrop trend, recognised in this thesis as an  $F_2$  fold trend, and hence the two are not directly related. He considered the metamorphism to be related to Devonian granitic mobilisation later than the folding. The suggested relationship fails to account for several features:

- (i) If the granitic intrusions at depth account for the regional metamorphism of rocks, the metamorphic effect should increase with proximity to the granitic bodies exposed to the east of the study area south and east of Mount Albert. In the study area, however the rocks do not increase in metamorphic grade eastward.
- (ii) Since the intrusions are Devonian in age, it is logical, as Mattinson assumed, that the metamorphism was also Devonian in age. However, an absence of regional metamorphic effects from Siluro-Devonian rocks near the granites or elsewhere precludes the possibility that such effects are related to the Devonian granitic masses.
- (iii) Mattinson also contended that the gravity in the Shickshock area (Tanner and Uffen 1960), which suggests a rise in the basement, is directly responsible for the regional metamorphism. But the gravity data is contradictory with granitic emplacement as a granitic body should give rise to a negative anomaly. Tanner and Uffen (1960, p.236) observed that the gravity positive over the Shickshock Mountains is

very unusual and suggests that they have no roots. The anomaly is probably due to the rise of Precambrian gneisses.

It is considered in the present thesis that the regional metamorphism is pre-Silurian in age and possibly pre-Lower Middle Ordovician. It may have accompanied intense folding ( $F_1$ ) or mobilisation of the rocks following the folding. An attempt to relate the isograds to the early fold trends was given up due to extensive destruction of the older structures by the later folds ( $F_2$ ). The following points are relevant to such a deduction.

(i) Middle Ordovician rocks that have consistently yielded graptolitic faunas are only slightly metamorphosed. An exposure immediately south of the Shickshock Fault (fig. 6) is a black mudstone (Appendix V). Slates exposed one mile south of Matane (de La Rue 1941) which also yield Middle Ordovician graptolites are scarcely metamorphosed.

(ii) Metamorphic segregation, which produces the layering parallel to the schistosity within the Shickshock meta-volcanic rocks (Mattinson 1964) are folded by Devonian (Acadian) folds. Hence metamorphism precedes the  $F_2$  folding (microphotos. 2,3, and 4).

(iii) The isograds drawn by Mattinson are oblique to the Middle Devonian (Acadian) structures imposed on the Shickshock Group of rocks. The isograds are probably not related to the Devonian deformation.

Table 3. CHARACTERS OF FOLD GENERATIONS

Deformation phase	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Geographic distribution	Belt of Cambro-Ord. rocks. Shickshock Mountain area	Throughout the area	Northern coastal area	Southern part of Cambro-Ordovician rock belt
Scale of occurrence	All scales, penetrative on outcrop scale	All scales, penetrative.	Macroscopic and mesoscopic, Non-penetrative	Mesoscopic and Microscopic, Non-penetrative
Foliation surface involved	Lithologic layering S <sub>s</sub> .	Bedding S <sub>s</sub> and Schistosity, S <sub>1</sub> .	S <sub>s</sub> and axial plane fracture cleavage, S <sub>2</sub>	S <sub>s</sub> and S <sub>1</sub>
FORM:				
Profile	Not clear, isoclinal.	Open to isoclinal; box shaped; rounded hinges; similar in shale, concentric in limestone and sandstone.	Open to tight, S-shaped, asymmetric, concentric.	Open, angular S- or Z shaped, concentric.
Axes	Locally rectilinear.	Curved.	Vary with attitude of previously folded S <sub>1</sub> and S <sub>2</sub> .	Rectilinear on mesoscopic scale.

	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Axial lineation	L <sub>1</sub> ; Intersection of S <sub>s</sub> and S <sub>1</sub> , fine striation and or mineral elongation.	L <sub>2</sub> ; Intersection of S <sub>s</sub> , S <sub>1</sub> , with S <sub>2</sub> , crenulation on S <sub>s</sub> and S <sub>1</sub> surfaces.	L <sub>3</sub> ; intersection of S <sub>s</sub> , S <sub>2</sub> , with S <sub>3</sub> , irregularly spaced and rare	L <sub>4</sub> ; fine angular corrugation, spacing variable on S <sub>s</sub> and S <sub>1</sub> .
Axial surface	Folded parallel to layering.	Planar, divergent upwards on regional scale.	Planar.	Planar.
Axial foliation.	S <sub>1</sub> , well developed Mica and chlorite alignment.	S <sub>2</sub> , well developed fracture-slip cleavage.	S <sub>3</sub> , rare rough irregular fracture resembles closely joints.	S <sub>4</sub> , rare and irregularly developed fractures.

#### ORIENTATION

Axes	Low plunge, SW-NE.	Variable plunge lying in a plane striking NE or ENE.	Plunge steep, 60-90°SE	Lie on S <sub>s</sub> or S <sub>1</sub> High angled pitch. NW-SE
Axial surface	Folded by F <sub>2</sub> , strike NE or ENE.	Strike NE. Inclination between 50°SE-60°NW.	Strike E-W and NW. Vertical or sub-vertical.	Strike NW. Vertical or sub-vertical.
Symmetry.	Not clear.	Monoclinic/triclinic.	Monoclinic.	Monoclinic/triclinic, sometimes symmetric.
Relationship of lineations to each other.		On small scale L <sub>1</sub> and L <sub>2</sub> rarely oblique. In stereo plots L <sub>1</sub> and L <sub>2</sub> co-axial	L <sub>2</sub> ∧L <sub>3</sub> high angle.	L <sub>1</sub> ∧L <sub>4</sub> and L <sub>2</sub> ∧L <sub>4</sub> high angle.

ORIENTATION	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Characters establishing relative age of structures		Folds affect S <sub>1</sub> . Crenulation lineation. L <sub>2</sub> obliterates L <sub>1</sub> .	Folds affect S <sub>2</sub> .	Folds affect S <sub>s</sub> , S <sub>1</sub> . Planar character of F <sub>4</sub> axial surface across S <sub>s</sub> and S <sub>1</sub> deformed by F <sub>2</sub> indicates F <sub>4</sub> later than F <sub>2</sub> . Age relationship between F <sub>3</sub> and F <sub>4</sub> not clear.

In the description, while indicating the relationship of various structures to the local fold geometry, fabric symmetry axes (Sander 1930) a, b, and c are used as defined below:

ac, the symmetry plane in monoclinic folds.

b, normal to ac plane.

a, perpendicular to b in the principal fabric plane (axial or cleavage plane which is thus ab).

c, normal to ab plane.

The axes are equivalent to A, B and C axes suggested as monoclinic fabric symmetry axes by Ramsay (1967, p. 335).

Double plunging folds do not always have a monoclinic symmetry and therefore orientation of the b axis can be confusing. The bedding surface in the very initial stages of folding can be assumed as horizontal and containing the b axis (Ramsay 1962b). Thus, the b axis in variably plunging folds is assumed as having remained horizontal or sub-horizontal in the ab plane. The double plunging folds also have similar fold relationships and they are considered as formed due to movement parallel to the kinematic a axis, which can be parallel to the geometric a axis in the axial plane; b axis is normal to the a axis in the axial plane.

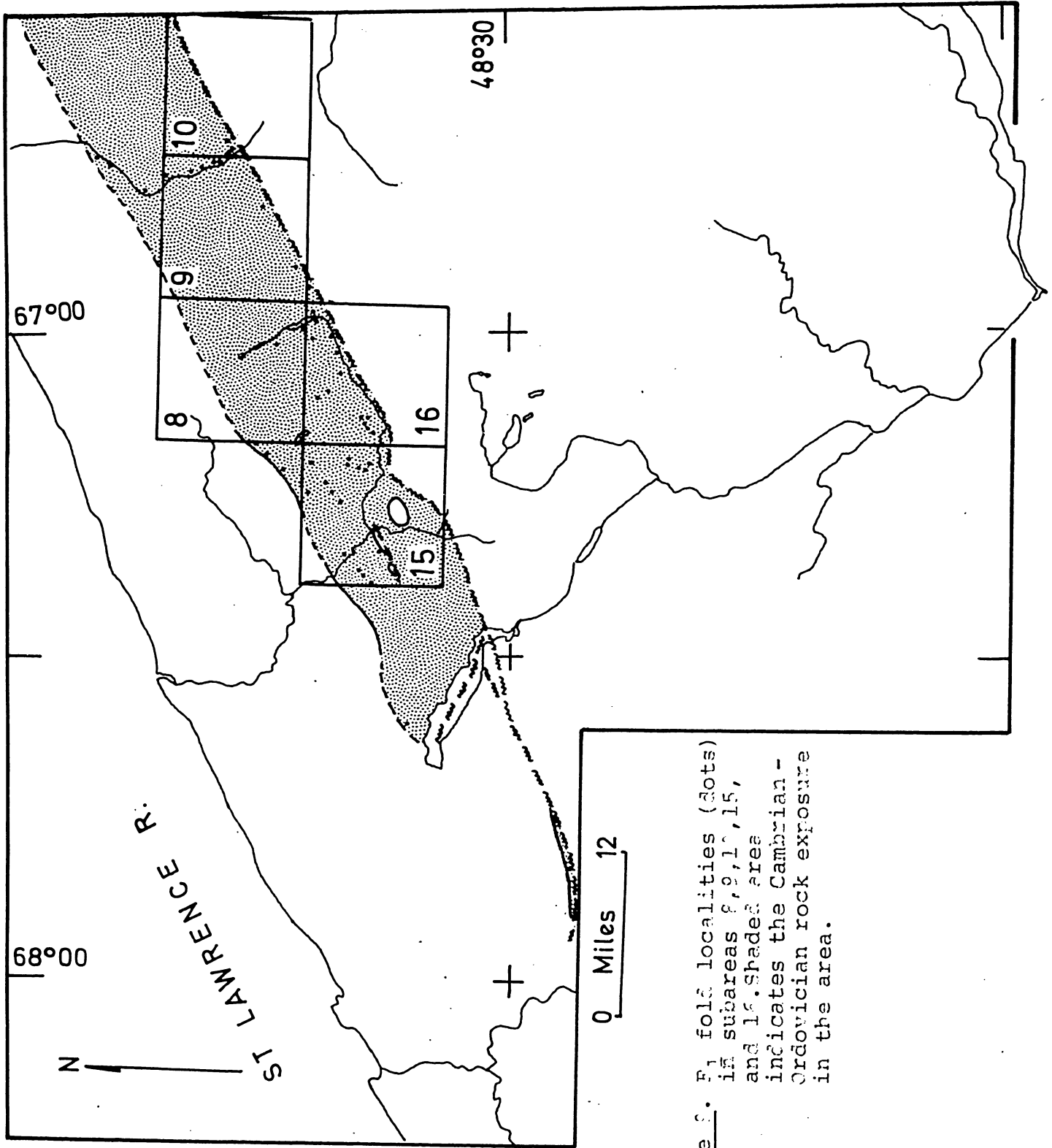


Figure 3. F<sub>1</sub> fold localities (dots) in subareas 8, 9, 10, 15, and 16. Shaded area indicates the Cambrian-Ordovician rock exposure in the area.

## III STRUCTURAL GEOLOGY

1. The  $F_1$  FoldsIntroduction

The  $F_1$  structures are restricted to a belt of Cambrian-Ordovician rocks, 6-10 miles wide in the Shickchock Mountain area, and a narrow inlier about 2 miles long and about 1500 feet wide southwest of Matapedia Lake (fig. 9). Metavolcanics and metasedimentary rocks including arkosic sandstone (Shickshock Group) and phyllitic slate are affected by  $F_1$ , and have been deformed by later folds. Consequently they contain multiple structural elements.

Profile

Since in almost all outcrops examined,  $S_1$  schistosity, an axial-plane foliation of  $F_1$ , maintains a close parallelism with the layering ( $S_g$ ) in the later folds ( $F_2$ ) which deforms both these structures, it is considered that they were isoclinal before refolding. In the Shickshock Group of rocks the schistosity is folded in the same manner as the bedding in younger rocks. This suggests that before folding by  $F_2$ , schistosity in the Cambrian-Ordovician rocks was parallel to the horizontal bedding in younger rocks and that the  $F_1$  folds were recumbent. Fold closures are rare due to poor exposure in the Shickshock Mountains. Two examples of the  $F_1$  folds were observed from the belt of grey slates exposed north of the Shickshock Mountains north of the Mount Albert east of the area (photo 1 and 2).

The lack of recognisable  $F_1$  folds is also due to the fact that in the field, folds in the metavolcanic rocks are mainly traceable by following secondary granular quartz veins. The veins are parallel to the schistosity  $S_1$  and are apparently a post  $S_1$  structure. Therefore in the field, one follows the quartz vein fold closures ( $F_2$  folds), and the  $F_1$  folds go undetected.

Planar and Linear Elements

The planar elements of the  $F_1$  folds consist of a schistosity  $S_1$ . It is characterised by recrystallised micas, hornblende, chlorite,

and sericite aligned parallel to  $S_1$  (microphotos 2, 8). The  $S_1$  schistosity is parallel to bedding  $S_g$  in most cases. The two structures diverge occasionally in the area of Cambrian-Ordovician slates north of the Shickshock Mountains. Even in very tight folds, suspected as  $F_1$  folds in the metavolcanics, the  $S_1$  schistosity was found parallel to  $S_g$  in every part of the fold (microphoto 2,3 and 4). These are therefore younger folds.

Arkosic metasandstone, exposed in a belt immediately north of the Shickshock Mountains, demonstrates well developed schistosity in the matrix (microphoto 1). It consists of recrystallised chlorite and muscovite. The schistosity is tectonic because: (i) The authigenic mica that defines  $S_1$  bends around quartz fragments. (ii) Authigenic plagioclase abuts against micas. (iii) Pre- or synkinematic plagioclase and quartz probably are rotated in relation to schistosity which is deflected near the porphyroblasts..

In the inlier of Cambrian-Ordovician rocks southwest of Matapedia Lake, the rocks are phyllitic. A pronounced schistosity ( $S_1$ ) is parallel to layering.

Sixty-five miles east of the area, near Mount Albert,  $S_1$  schistosity appears to bend around the ultrabasic body (fig. 29) (Appendix III).  $S_1$  is well developed and has a characteristic silvery lustre. It is cut obliquely by a steep slip cleavage,  $S_2$  (exposed 200 yards upstream along Ste. Anne Northeast Stream before it joins Ste. Anne River under the old bridge).

Throughout the study area the axial linear elements  $L_1$  of the early folds trend northeast and are characterised by fine striations, and where not parallel, by  $S_g$ ,  $S_1$  intersections.  $L_1$  is also formed by elliptical lensoids of cleaved dark green chlorite, up to one inch long, lying on the schistosity surface (photos 4 and 5), which are presumably elongated parallel to the  $F_1$  fold axes. These are low pitching and accompanied by a fine striation parallel to the long axes of the ellipses. Boudins in locality RM-1 (photo 3) may have formed during  $F_1$  folding as they are parallel to the  $F_1$  fold axes (cf. de Sitter, 1958).

Professor E. Cloos (verbal communication 1967) mentioned chlorite 'blebs' in the Precambrian gneisses along the Skyline Drive between Potomac and James Rivers in Maryland. The blebs are much

larger than those observed in the Shickshock Mountain area, and pitch down the dip of the gneisses. He regarded them as a lineation in "a" (Cloos 1946, p.25) in the principal fabric plane (Sander 1930, p. 119) as they conform to the direction of the steep slickensides on calcite veins oblique to the axial planes of the local folds.

In the study area however, the chlorite ellipses are considered as structural features of an origin different from that of the Skyline Drive in view of the following: (i) The long axes of the chlorite ellipses in the study area are subhorizontal. (ii) Lack of slickensides or grooving parallel to the long axes of the ellipses. (iii) The chlorite ellipses are presumably parallel to the  $F_1$  fold axes.

In the Shickshock Group a striation  $L_1$  is occasionally observed oblique to the later  $F_2$  fold axes. In one example (photo 7) the angle of divergence is at a maximum along the hinge and smallest along the fold limbs, but the deviation is too small to warrant deductions about the deformation mechanism as deduced by Ramsay (1960).

A similar divergence between  $L_1$  and  $L_2$  is observed in grey slates in locality RM-1, near Riviere Matane village, where a faint striation is observed oblique to  $F_2$  axes. The striation is obliterated along the hinge area, where extensive  $L_2$  develops as a crenulation.

Thus,  $L_1$  is locally oblique to  $L_2$ . However on a large scale, as demonstrated in a stereo plot,  $L_1$  and  $L_2$  coincide on an average, and are difficult to distinguish. Thus it is practically impossible to subdivide the area into domains with respect to consistent  $L_1/L_2$  angular relationship.

#### Age of the $F_1$ Folds

Several workers have indicated a pre-Upper Ordovician deformation in the western Gaspé area. (de LaRue 1941, McGerrigle 1954, Ollerenshaw 1963). The presence of an Upper Cambrian-Lower Ordovician deformation is justified in view of the following:

(i)  $F_1$  structures, the schistosity  $S_1$  and lineation  $L_1$ , are restricted to the Shickshock Group and Cambrian-Ordovician metasediments to the north. Lower Ordovician and younger rocks of

similar lithology further north do not contain  $S_1$ ; for example:

(a) Arkosic sandstone structurally underlying the Shickshock Group and included by Ollerenshaw (1963) in the group, is exposed along the northern margin of the Shickshock Mountains as a thin band, south of Duvivier Lake and north of Matane Lake. It has a strong  $S_1$  schistosity of tectonic origin in the micaceous matrix (see p.99) (microphoto 1). The fabric can be compared with that of impure grey sandstone northeast of Ste. Felicite on the St. Lawrence Coast (the stratigraphic equivalent of the Pillar Sandstone of Early Ordovician age, McGerrigle 1954. (microphoto 20 and 21). The absence of the schistose fabric in the Lower Ordovician sandstone indicates that the deformation may possibly be older. However, the two localities are 17 miles apart, and, the absence of the  $F_1$  structures from the sandstone could also be due to geographic reasons. The tectonic front of the  $F_1$  folding may not extend to the Ste. Felicite area.

(b) Ollerenshaw (1963) reported an outlier of sandstone in Cambrian-Ordovician rocks along the Tamagodi Stream a mile upstream from the confluence of the Cajetan Stream, 6 miles southwest of the Shickshock Mountains (location shown in fig. 6). This sandstone resembles the impure grey sandstone and Pillar Sandstone and probably has the same Lower Ordovician age. The sandstone is poorly exposed and unconformable to the underlying rocks. It is regarded as derived by erosion of the Shickshock volcanic rocks. The fabric of the sandstone with an absence of schistosity is similar to that of the impure grey sandstone.

(c) A wedge of graptolitic shale exposed along the Shickshock fault zone in the Cap Chat River valley (fig. 6) is dated as Middle Ordovician (Caradocian, Appendix V). These, along with other Middle Ordovician rocks, the age of which has been determined paleontologically by de LaRue (1941) and McGerrigle (1954), do not contain  $F_1$  structures.

(ii) Age of rocks deformed by  $F_1$ . The pre-Lower Ordovician rocks have not been precisely dated. However, a rare fossil find northwest of Matane Lake (fig. 6) from the belt of metasedimentary rocks

$\frac{1}{2}$  mile to the north of the Shickshock Mountains suggests a Middle-Late Cambrian age (Ollerenshaw 1963, See Appendix I, Age of the Cambrian-Ordovician Slates).

(iii) An ultrabasic igneous body, consisting chiefly of serpentinite, is intruded into the Cambrian-Ordovician phyllitic slate inlier southwest of Matapedia Lake. The contact zone is extremely metamorphosed (Beland 1960). The intrusive is presumably equivalent in age to the Mount Albert intrusive (see fig. 29 for location of Mount Albert igneous body).

The contact metamorphosed  $S_1$  schistosity in the metavolcanics along the aureole of the Mount Albert ultrabasic intrusive is dated as Lower Ordovician (MacGregor 1962). The deformation of the Shickshock Group rocks and the development of the  $S_1$  is dated radiogenically as Middle-Late Cambrian (Ollerenshaw 1963).

In view of the facts given, the  $F_1$  deformation, is regarded as pre-Middle Ordovician and probably pre-Lower Ordovician in age.

## 2. The F<sub>2</sub> Folds

### Introduction

F<sub>2</sub> folds affect all rocks throughout the area older than upper Middle Devonian. The trend of the folds varies from north-east to eastnortheast.

In profile the folds vary from isoclinal to open; the shape and size largely depends on the lithology of the deformed rock. In most cases the fine clastic rocks are finely plicated, especially in thin interbeds. The coarse massive calcareous and arenaceous units are broadly folded on macroscopic and regional scales.

Folds are best demonstrated where there are well defined marker horizons or interbeds, and they were therefore studied in detail in such rock types.

### Profile

Several features about the profiles of the F<sub>2</sub> folds in various stratigraphic units in the Lower Paleozoic succession are apparent from net diagrams. The acute and isoclinal nature of the folds and steepness of the limbs can be estimated from the single concentration of the polar plot of bedding ( $s_g$ ) in most diagrams (Plates III).

In the Shickshock Group the S<sub>1</sub> schistosity is folded symmetrically (microphoto 4) to asymmetrically (microphoto 2), open to isoclinal (photo 21, 22 and microphoto 2-3). The folds have rounded hinges.

Within the belt of Cambrian-Ordovician slates north of the Shickshock Mountains the S<sub>1</sub> schistosity defines F<sub>2</sub> folds that vary in tightness.

Owing to good exposure, the fold form is best studied within the Ordovician rocks along the coast. The folds range from open and conjugate (photo 37) to isoclinal (photo 36). The shorter limbs of the competent layers in the asymmetrical folds are sheared or thinned (Plates IIb3 and IIe5).

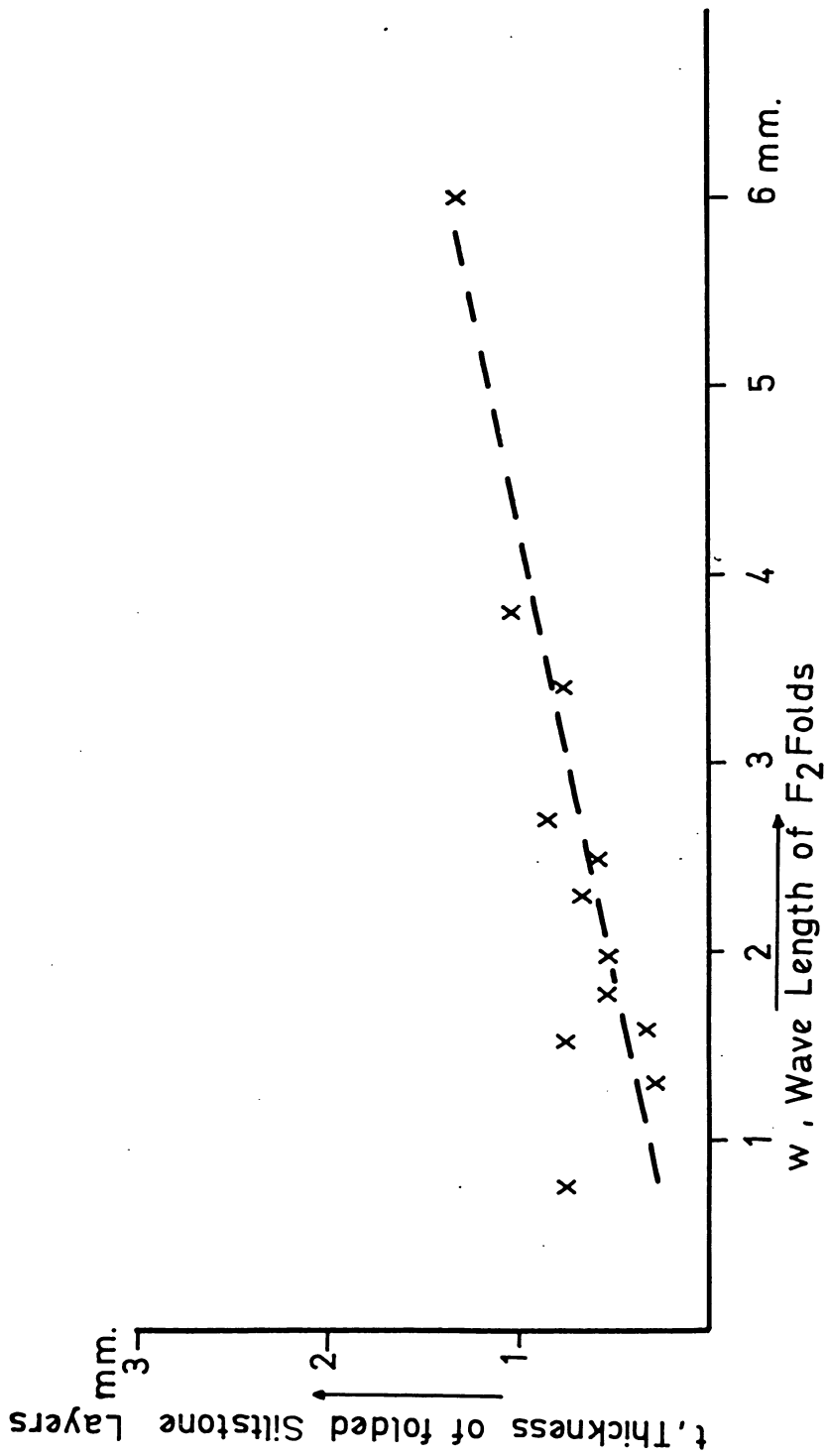


FIGURE 10. THICKNESS OF CALCAREOUS SILTSTONE LAYERS IN ORDOVICIAN SLATE, PLOTTED AGAINST F<sub>2</sub> FOLD WAVE LENGTHS ON MICROSCOPIC SCALE. SLOPE OF THE CURVE IS BETWEEN 1/4 AND 1/4.5 INDICATING A t/w RATIO OF 1:4-4.5 (12 FOLDS).

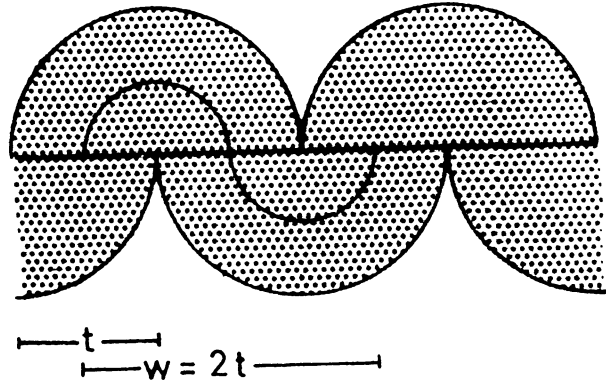


Figure 11. Diagrammatic maximum shortening of concentrically folded layer. The wave length is 2 times the thickness ( drawn after Ramsay 1967, p.387).

The  $F_2$  folds in the Silurian rocks, which consist of massive quartzite, limestone, and sandstone with minor shale are generally open to moderately tight with obtuse to acute interlimb angles. The folds are exposed on a regional scale near Amqui and Causapschal (fig. 6), but are lacking on smaller scale. The folds are tight to open in the Cape Bon Ami Formation (exposed along the railroad track north of Causapschal railway station), and open in the Devonian sandstones (exposed near Casault Lake).

In the Fortin Group, Lower and/or Middle Devonian in age, which consists of argillites of variable lithology, folds are varied in style. In massive calcareous slates they are open (photo 50) to moderately tight, and in micaceous shales they are acute to isoclinal (photo 51).

To allow a study of the mechanism of folding, the shape and style of profile of the folds were studied in detail. Small folds were sectioned and studied under the microscope. The larger folds were photographed looking down the plunge. These photographs were projected and accurately traced (Plate II).

The  $F_2$  folds in the Lower Paleozoic rocks are variable in wave length. The ratio of wave length to average thickness measured on the hinge and the limb areas of the folded competent unit (which usually consists of limestone or siltstone) deformed within a less competent (shale) bed, is also variable. For example in 12 small folds measured it is between 1.02 and 5.00. However, a scatter about a ratio of about 4-4.5 in the examples is suggested graphically (fig. 10).

It can be demonstrated geometrically (Ramsay 1967, p. 351) that in an isoclinally folded series of rocks there is a median surface along which perfect concentric relationships can be established for a particular layer so that there is no flattening (fig. 11). Above and below this surface squeezing occurs within other layers in the interiors of the folds. The wave length of the folds as measured along the median surface is at a minimum of 2 times the thickness of the layer (Fig. 11). This relationship involves the maximum possible shortening in a layer by concentric folding (shortening of 36%, de Sitter 1964, p. 277). It does not take into account the squeezing and flow of adjacent layers.

The ratio of 1:4-4.5 as measured in competent layers (fig. 10) indicates a shortening of the folds to about one half the maximum extent possible by perfectly concentric folding. However, above and below the layer squeezing and flow of enclosing (incompetent) material occurs parallel to the axial plane of the folds.

The hinges of successive competent layers separated by incompetent shale do not lie in a plane but are shifted towards the shorter limb of anticlines (photo 45). Therefore, the plane bisecting the limb angle for a competent layer is different from that of the adjacent competent layer. Thus, the axial plane is bent in a series of competent and incompetent beds during folding.

Where two competent layers are separated by a shale layer, the thickness of the shale layer is less in the longer limb than in the shorter limb of a fold (photo 45).

The variation in thickness of the competent layers in the folds were studied for 24 folds. On the diagrams in plates II. the thickness of a competent units was obtained at the hinge and along the limbs by measuring the perpendicular distance,  $t$ , between tangents on lower and upper surfaces of the units at the same angle to the axial plane of the fold (Ramsay 1962b). These thicknesses were taken at various distances along the beds as shown by the reference lines on each diagram. Different folded layers transect a reference line at approximately the same angle.

Similarly the thickness of both competent and incompetent layers,  $T$ , were measured parallel to the axial planes and plotted against the reference lines as in plates II.

The following observations were made on the thickness relationships of the siltstone and calcareous layers.

(i) Changes in thickness of the competent layers are often variable (Plate II). The competent layers are frequently cross bedded and indicate a fast current regime of deposition. In this regime a variation in the thickness of sedimentary layers is considered natural. Although the variation in the thickness is largely sedimentary, a tectonic influence is also apparent as shown by the following.

(ii) In many folds the competent layer is relatively thickened in the hinge areas and thinned on the limbs (plates IIb,3;

IIc, 2 and IIe, 5).

(iii) The competent beds with minor shale intercalations maintain a fairly consistent thickness (normal to bedding) across the folds conforming to a concentric model (plates IIa; 1, 2, 3, 4; IIb, 4, 5).

(iv) The comparison of thicknesses parallel to the axial planes demonstrate that in many examples, although the competent beds may vary, the shale layers have a fairly constant thickness across the fold profiles (plate IIa, 4; IIb, 4; IIc, 1, 5; IIe, 3). Although the thicknesses are not exactly constant they are close enough to indicate similar fold styles in the shale.

(v) The shale layers in some examples show a greater thickness, parallel to the axial planes, along the hinges of folds (plate IIb, 2; IIc, 2, 4).

A good example of a concentric fold is shown in photograph 94. The folded rock consists of limestone layers with minor shale intercalations. Bedding-plane slip is inferred from displacement of calcite veins that lie parallel to the fold axis and across the beds. The calcite veins are displaced across the thin shale layers. In the anticline shown the overlying beds are displaced upward toward the hinge of the fold, and the amount of displacement apparently decreases towards the hinge. In places where there is fine interbedding of shale and limestone the veins are displaced in a series of small offsets so that the vein itself is bent towards the fold hinge.

The calcite veins are not regularly oriented with respect to the fold geometry. It is difficult to visualise at what stage the calcite injection occurred into the structure. Consequently it is impossible to estimate the amount of movement purely by a concentric folding mechanism.

In a folded series of calcareous slates, the outer part of each folded layer of constant thickness separated by shale laminae, is tensionally fractured along the curvature of the folds (photo 42).

The following conclusions are made from the study of the profiles:

(i) In many folds (Plate IIe) the thickness of the competent

layers in proportion to the more ductile shale is small. Thus it appears unlikely that the competent layers transmitted compression throughout the development of the folds.

(ii) The thickness relationships do not conform ideally to the concentric or similar models.

(iii) The thinning of the limbs of folds in competent layers indicates that the shortening of folds was accompanied by flattening of these layers, and where the shales are constant in thickness parallel to the axial planes similar folding was operative.

(iv) The fairly consistent ratio of the thickness of the competent layers to wave lengths (fig. 10) is not random, as it would be by similar folding wholly due to unequal movement ('flow') along the axial planes. A buckling element is therefore considered to have determined the wave lengths. It is suggested that the folds may have been initiated by buckling and flexure slip between competent beds in the early phases of deformation. The competent layers were active members (Donath and Parker 1964).

(v) The succeeding deformation was dominated by similar folding within the shale layers which became the active constituents, and the competent layers were deformed passively. This gave rise to folds in the incompetent layers that commonly approach the similar type. The competent layers were slightly modified in thickness. The folds thus formed probably had their wave lengths reduced from the original determined by buckling. Rare thinning of the competent layers at the hinge of the folds is due to fracturing of the competent layer and injection of shale into the fractures (photo 31). In other cases the thinning is sedimentary (photo 44, drawn in plate IIc, 5).

An extension in the curved competent layer, usually along the limb areas and rarely along the fold hinges is considered as due to flow of the incompetent layers and extension of the competent 'caps'. The pushing of the 'caps' of competent layers outward from the cores of the folds and breaking apart of the competent layers resemble the behavior of flexible

sheet which accomodates to the compression in the interior of a fold.

According to Ramsay (1962b, p. 313) the ratio of the thickness of a layer in the hinge area to its thickness on the limb (measured so that its angle to the axial plane between parallel tangents to the surfaces of the layer) gives the degree of flattening. In the competent layers illustrated the maximum ratio is .55 in plate IIb, 1, and accordingly, the percentage flattening within the competent layer is approximately 30-33% across the axial plane (Ramsay 1962b, p. 315).

Shortening across the axial planes gave rise by a continuous process to the near concentric and similar folds. The similar folds resulted from a continued shortening due to compression, and an advance in the normal direction along axial plane fractures, or in absence of fractures, by plastic deformation and flow. In most cases a combination of both occurs. Between fractures the microlithons (de Sitter 1964, p. 268) are plastically deformed (see nature of the  $S_2$  fracture cleavage).

### Geometrical Elements

The axial linear elements  $L_2$  of the  $F_2$  folds consist of fold axes, bedding  $S_g/S_2$  cleavage intersections in the rocks which are not deformed by the  $F_1$  folding, and schistosity  $S_1/S_2$  cleavage intersections in the rocks that are deformed by the early phase of folding. A crenulation of  $S_g$  or  $S_1$  develops as an expression of the axial lineation, but is usually restricted to the hinge areas of  $F_2$  folds.

The lineation  $L_2$  lies on the mean axial plane  $S_2$  (or the mean axial plane cleavage) at variable angles (Plate IV a, b). Where the fold hinges were not exposed as in the interior, this method for deducing the axial plane is most useful, and consists of plotting  $L_2$  on a stereonet. The great circle that passes through the  $L_2$  plot is the mean of the axial plane of the  $F_2$  folds.

The trace of the  $S_2$  cleavage on bedding surface is not regular. It often consists, in detail, of non-parallel lineations, linear ridges or traces of fractures, that give rise to lenticular segments.

Occasionally the upper bedding surfaces in anticlines show a bulge between cleavage traces (photo 56, discussed on page 45).

The  $L_2$  lineation on  $S_2$  cleavage surfaces is marked by the trace of a lithological or colour variation.

From the stereographic plots (plate IVb) it is apparent that the average orientation of  $L_2$  and that of the meso-fold axes as directly measured are coincident. The average plunge is parallel with that of the  $F_2$  folds on the regional scale.

Fold Axes. The  $F_2$  fold axes vary in plunge from horizontal to vertical and in pitch from  $0^\circ$  to  $90^\circ$ . They plunge in either NE or SW directions or are double plunging.

In several subareas more than one maximum of fold plunge is developed (fig. 14). This demonstrates the variation of fold plunge in each subarea.

The double plunging folds are exposed on all scales, and their outcrop pattern resembles a canoe (photo 47). The double plunging nature of folds is ascribed to a single phase of folding because of the following observations:

- (i) Folds plunging NE occur side by side with those plunging SW and no alignment of crests or troughs is apparent.
- (ii) Axis of folds commonly curve from a horizontal to steep inclination while the adjacent folds maintain a constant plunge.
- (iii) Folding of the planar axial surfaces by later folds is not common and is restricted to the St. Lawrence coastal areas; This is demonstrated by the consistent trend in the axial plane fracture cleavage in the subareas (plate III).

$F_2$  Axial Planes. In well exposed outcrops the axial surfaces of the folds were directly measured. However, owing to poor exposure of rocks in the interior, the number of direct measurements on axial surfaces of  $F_2$  folds is not significant. In these the plot of the axial plane cleavage  $S_2$  is used to indicate the axial planes. It is also deduced using the linear elements (see page 39).

The planar or semi-planar axial surfaces of the  $F_2$  folds of mesoscopic scale correspond in attitude to the axial planes of the

large scale folds.

In domain I and also to some extent in II (as defined on page 54) successively higher points on an anticline lie further to the northwest than if on a plane, so that the axial surfaces of the  $F_2$  folds are usually slightly curved, concave northward.

### Axial Plane Cleavage

Fanning of  $S_2$  cleavage is common in individual folds. In shaly rock it diverges upwards in synclines (photo 53 and 54), and converges upwards in anticlines (photo 44).  $S_2$  cleavage diverging upwards in an anticline is rare. On the scale of an outcrop on a stereonet the  $S_2$  cleavage and axial planes are coincident. The same is observed for the subareas (Plate IVb).

The cleavage is essentially a fracture set parallel to ab axes (Sander 1930). It can be defined according to different characters as fracture cleavage (photos 55 and 64), slip cleavage (microphoto 15) and crenulation cleavage (photo 57). Each of these examples would define a specific feature of a variety of development of cleavage in the area. Furthermore, along the limbs of tight folds where the cleavage parallels bedding it is associated with partial recrystallisation of micas and sericite (as observed in Fortin Slate near Ste. Florence). Thus, it could be also defined as partly a slaty cleavage.

In general it consists of a fracture or a zone of fractures which bends around quartz rich zones, blocks of limestone, silty layers or clastic grains within shaly rock (microphoto 15, 17; photo 99). The term axial plane cleavage is adopted because it defines a geometrical relationship and general character and does not imply a mechanism or a kinematic feature.

A common feature among different developments of the axial plane cleavage is that each cleavage plane is a zone of displacement of the surfaces ( $S_g$ ) crossed. The displacement in rocks varies from macroscopic to submicroscopic in scale according to the rock type and degree of folding. Examination under the microscope reveals that the displacements may either be marked by a clear cut break (microphoto 8 and 9), or may consist of an extension and rotation of the reference horizon without rupturing (microphoto 15). A fracture may displace

a thin bed, but against a thicker competent layer the same fracture may result in a flexure without rupture (microphoto 10).

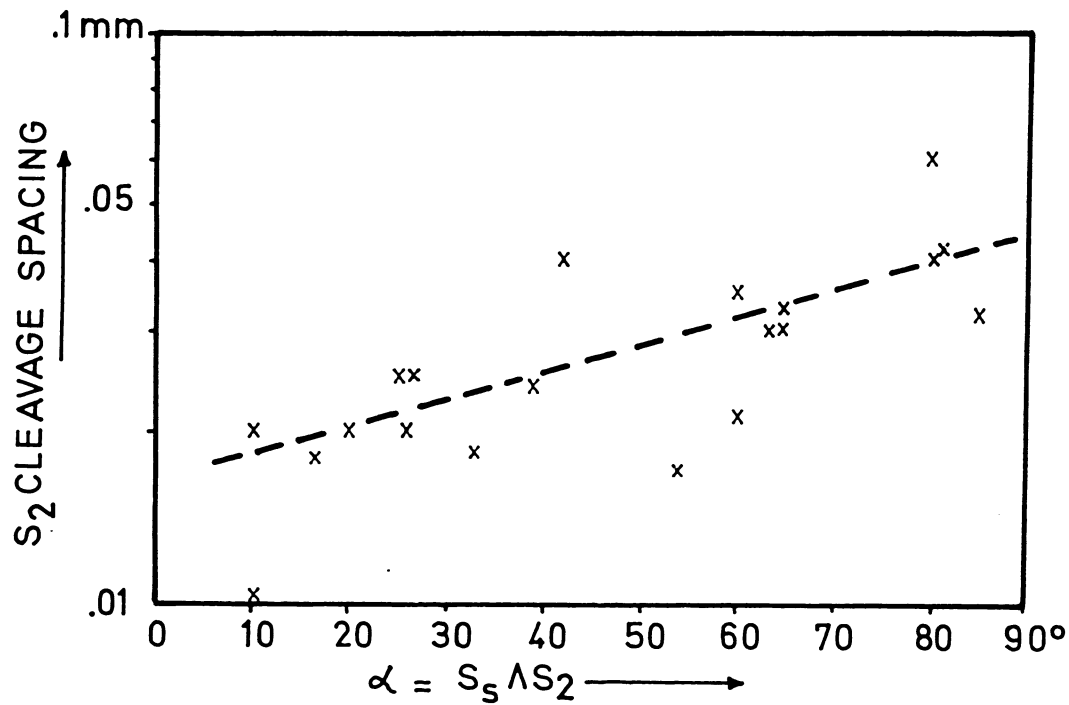
Each cleavage plane displaces the reference horizon. The regularity of the amount of displacement across the cleavage depends on the lithology. When the reference horizon within shale consists of only colouration contrasts each cleavage plane displaces the horizon varying small amounts (microphoto 13). If the reference layer is relatively competent and thick, it is displaced relatively large distances across a few of the planes of cleavage penetrating the adjacent incompetent layers (microphoto 9). This results in breaking up of the competent layer into slabs, which may be pushed outward from the cores of the folds (photo 99). This punching out of the competent beds into the finer lithologies is often observed macroscopically (photo 54).

In fine uniform lithologies the trace of the cleavage on the ac surface is linear and traces are parallel to one another. Traces are irregular and wavy in non-uniform lithologies. On bedding surfaces the traces of the cleavage fractures parallel one another or merge in short or long lens shaped segments depending on lithology.

Discussion on the Definition of Fracture Cleavage. Fracture cleavage has been regarded by Leith (1923) as a closely spaced variety of joints. Whitten (1966, p. 260) rejects fracture cleavage as a variety of axial plane and crenulation foliation because it is not determined by mineral orientation and is non-penetrative.

The worker disagrees with Whitten's exclusion for the following reasons:

- (i) Axial-plane foliation according to definition should include all cleavage or foliation parallel to the axial planes of folds. The criterion is exclusively a geometrical relationship.
- (ii) Crenulation cleavage is only partly due to mineral re-orientation by cleavage development. It is a crenulation of a pre-existing anisotropy, and in many cases, the transection by a set of cleavage fractures.
- (iii) Whether the cleavage is considered as penetrative or not is subjective. It is essential to mention the scale



**FIGURE 12.** SPACING OF S<sub>2</sub> CLEAVAGE PLANES PLOTTED AGAINST  $\alpha = S_5 \wedge S_2$  IN ORDOVICIAN SLATE. THE GRAPHS DEMONSTRATE AN INCREASE IN SPACING WITH GREATER  $\alpha$  IN FOLDS WITHIN SAME LAYER (20 OBSERVATIONS IN 4 FOLDS).

while referring to the penetrative character of the cleavage.

In the area the fracture cleavage which is subparallel to the axial planes of folds, is a synkinematic and kinematically active structure. The fracture cleavage also separates microlithons plastically deformed and extended parallel to cleavage fractures. Joints on the other hand are post deformation structures. Therefore, the cleavage and joints, although both sets of fractures, are treated as distinctly different structures.

S<sub>2</sub> Cleavage and Degree of Folding. The spacing of the S<sub>2</sub> planes depends on lithology and degree of folding. The effect of degree of folding can be studied in a particular layer within which the lithology is more or less uniform. In the field the cleavage frequency is observed to be inversely related to the angle between the S<sub>g</sub> and the S<sub>2</sub> surfaces. This crude field observation was tested more precisely on the microscopic scale. The spacing of cleavage planes was measured at different parts of the folds. With each observation the angle between the S<sub>2</sub> and S<sub>g</sub> ( $\alpha$ ) was measured in the same lithology. Measurements on the spacing were plotted against  $\alpha$  (fig. 12). It is seen that the greater the angle of intersection  $\alpha$ , the wider is the spacing between the cleavage planes; that is, the tighter the fold, the greater the frequency of the cleavage. It also implies that the cleavage is more frequent along the limbs of the folds than along the hinge. In a fold limb, a flexure giving rise to a decrease in the angle  $\alpha$  results in an increased development of cleavage (microphoto 16). In microphoto 6 a greater frequency of S<sub>2</sub> cleavage along the limbs than at the hinge area is well demonstrated. In a series of tight folds in microphoto 7 the hinge areas continue as zones of no or infrequent cleavage.

Relationship of the S<sub>2</sub> Spacing to Lithology. In coarse and competent rock types the folds are generally open and the spacing of the fracture is widely apart. The displacement across the plane of fracture is either extremely minute or absent. A reason for the lack of apparent displacement could be that small scale displacements are masked by the absence of thin reference horizons in coarse lithologies.

In fine incompetent rock types, shale, mudstone and laminated shale with silt and limestone, the axial plane cleavage  $S_2$  demonstrates a great variation in the nature of the fracture and also spacing. The main difference, however, arises from either (i) the absence or varying amounts of displacement of the folded layers across the cleavage fracture, or (ii) the degree of anisotropy parallel to the folded surface. The displacement across the  $S_2$  cleavage is important because it distinguishes between the definitions of fracture and slip cleavages.

Spacing of the cleavage is variable. In open or close folds (Fleuty 1964) it is .02 mm in sericitic micaceous shale. The spacing increases through an increase in competency of lithology: In slightly calcareous shale the average spacing is between .5 to 1mm and more: In calcareous shale and shaly limestone or arenaceous beds the average spacing is 1-3 cms or more (photos 62, 63 and 64).

The  $S_2$  fracture is poorly developed in the Shickshock Group, commonly it is absent. It is present as an irregular microscopic fracture cleavage in the grey slates of Cambrian-Ordovician age. It is an irregular fracture depending on degree of foldings variably spaced at .3 cm to 1.5 cm in the Ordovician slates. In the shaly limestone of Cape Bon Ami Formation of Devonian age the spacing is .3 cm to 4 cms. In the Devonian sandstone the spacing is approximately 5 cms, and in the Devonian slates the spacing is variable according to the lithology and degree of folding.

In coarse sandstone and greywacks the cleavage often resembles a set of strike joint sets. In the field the difference between the two is as follows:

- (i) The cleavage has various angular relationship to bedding. The strike joints are usually normal to layering.
- (ii) The  $S_2$  cleavage fracture is regular in spacing and development.

The cleavage fractures abut against the competent limestone or calcareous siltstone layers and occasionally traverse them. This depends on the degree of development of the cleavage and lithological contrast. Refraction of cleavage is common and the angle  $\alpha(S_2/S_g$  or  $S_2/S_1)$  increases as the cleavage traverses the more competent layer.

A competent layer (calcareous siltstone or limestone) within shale when traversed by the cleavage usually exhibits a series of features which accompany folds:

(i) An increase in the frequency of cleavage planes and convergence towards openings on both sides of the layers takes place (photo 99).

(ii) The detached slab or the block of the competent layer is usually moved or rotated. The cleavage bends around such competent 'inclusions' (microphoto 15).

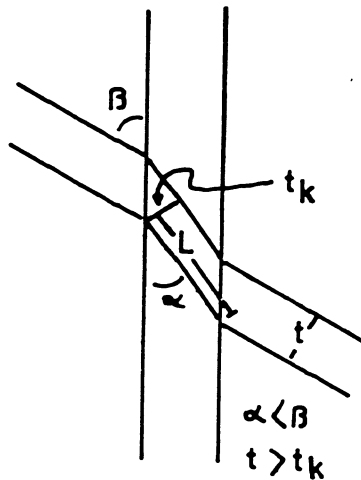
The convergence and the increase in the frequency of the cleavage appears to follow the direction which allows a greater 'flow'. The opening in the 'cap' of competent layer appears to resemble an opening for a confined fluid. A restricted opening makes the flow lines converge towards the 'outlet'. The angle of the cleavage on such instances is oblique to the general direction in a fold, but across the restriction it diverges back to the normal attitude, parallel or subparallel to the axial plane of the fold.

The convergence of the cleavage through the openings between the slabs is partly due to concentration in the shortening.

Deformation Between Fractures. Fracture cleavage with crenulated microlithons is occasionally observed in the area (photo 57). In such occurrences the fracture joins the limb areas of a series of folded surfaces. It is observed in rocks with an inherent closely spaced anisotropy parallel to the folded surface. This anisotropy is an early schistose fabric ( $S_1$  in the Cambrian-Ordovician metavolcanic, meta-sedimentary rocks and phyllitic slates) or fine sedimentary interbedding of layers of contrasting competency in the Ordovician and younger rocks. The layers between the fractures are deformed by folding, which may include both buckling and flattening.

The rocks which do not contain closely spaced planes of anisotropy parallel to layering also demonstrate such a 'plastic' deformation within the microlithons. This is indicated by a bulge between the  $S_2$  fracture trace on bedding surfaces (photo 56). This probably occurs by a shortening of the distance between two cleavage fractures, and a plastic extension of the microlithon normal to it.

Cleavage fracture  
under high  
magnification



.01mm  
approx.

Figure 13. Diagrammatic features in cleavage fracture 'zone'. It is assumed that (i)  $L$  extends in cleavage 'zone'. (ii) Thickness  $t$  decreases ( $t > t_k$ ). (iii) The width of cleavage continuously decreases. (iv) Angle  $\alpha$  between rotated foliation and the cleavage 'zone' decreases.

The crumpling of the microlithon is not uniform in spacing throughout a fold. Occasional single crenulated microlithons also occur, usually restricted to the hinge area of  $F_2$  folds (photo 57). The crenulations are often separated by slip surfaces.

Although at low magnifications the cleavage appears as clean cut planar surfaces, under high power objective each fracture is irregular and consists of a series of fractures. The lithological layers often continue across the cleavage 'zone', characterised by microscopic and submicroscopic sets of fractures. This is to some extent, though not clearly, shown by microphoto 11.

In the zone of cleavage (fig. 13) the rotated microlithon is thinned ( $t > t_k$ ) and the angle  $\alpha$  between the rotated foliation and cleavage zone is less than  $90^\circ$ . Assuming planar strain, as  $\alpha$  becomes smaller with the development of cleavage the reduction of  $t_k$  would imply an increase of  $L$ .

The shortening normal to the cleavage plane and deformation of the microlithon is considered a continuous process which results in the development of the  $F_2$  folds and the cleavage.

Classification of the  $S_2$  Fracture Cleavage. The spacing of the axial plane fracture cleavage planes is not significant in itself for classification. For instance, in two examples of equal spacing, one may show no slip, and the other may show sufficient displacement to be classed as a fault. Therefore a better means of classifying the fracture cleavage would be to use the slip to spacing ratio,  $d/D$  ( $d =$  average displacement across the cleavage surface, and  $D =$  average spacing of the cleavage).

In calcareous slate and shaly limestone the cleavage consists of widely spaced fractures with extremely small displacements. A figure of .05 mm or less for the displacement is considered average. With an average spacing of say 2 cms. the  $d/D$  ratio is .025 and the cleavage is called a fracture cleavage (with no significant slip).

In shale a cleavage spacing of .05 mm is common. The displacement is often .05 mm and even 2-3 times greater than the spacing. A ratio of 1 or higher is common for  $d/D$ . The displacement is significant, the cleavage is called slip cleavage.

An arbitrary ratio can be imposed to distinguish between fracture and slip types of axial plane cleavage, the essential difference between the two being the degree of displacement across the cleavage planes. A figure of  $\frac{1}{4}$  (.25), for  $d/D$  is taken as a convenient defining ratio; above this a cleavage is called slip and below called fracture.

Since the frequency of cleavage is directly related to its angle to layering, the cleavage can be of different types in the hinge and limb areas within the same fold. It is also observed that the slip across the cleavage surface is greater in tight folds than in open folds, therefore, the nature of the axial plane cleavage also differs with the tightness of the  $F_2$  folds.

The axial plane cleavage is further divided into the categories shown in table 4.

Table 4. Classification of  $S_2$  Axial Plane Cleavage in the Area.

	Mineral alignment	Slip-Spacing ratio $d/D$	Marker surfaces in microlithon	Smallest Penetrative Domain
Slaty Cleavage	Parallel orientation of mineral grains			Microscopic or Sub-microscopic
Fracture Cleavage	Lack of parallel mineral orientation. Defined by clean cut fracture or (under magnification) a zone of fractures	(Non-Slip) Fracture Cleavage $d/D < .25$	Surfaces unfolded or absent. Simple Fracture	Macroscopic or Microscopic
		Slip Cleavage $d/D > .25$	Surfaces unfolded, Simple Slip <hr/> Surfaces folded. Crenulation Slip.	

#### Calcite and Quartz Veins

Because of good exposures of the Ordovician rocks along the coast of the St. Lawrence River and poor exposure elsewhere, most of

the observations on calcite veins are restricted to this area, where they are related to  $F_2$  structures.

The veins in the Cambrian and Ordovician slate, limestone, and coarse clastic rocks consist essentially of calcite and rarely quartz. The Ordovician slates have abundant calcite veins. Usually they are less than 0.4 inches thick, but veins as thick as two feet are not uncommon. In general, the veins in the Shickshock metavolcanic rocks consist of quartz, epidote lensoids or calcite-quartz parallel to dip joint sets. In the same rocks secondary quartz and calcite layers parallel to the  $S_1$  schistosity are refolded. These appear to have intruded along  $S_1$ .

In the Siluro-Devonian slates and argillaceous limestone the veins are calcitic.

The most common variety of calcite veins is the ac vein (parallel to the a and c axes of the folds). Some oblique calcite veins are displaced horizontally up to 1 inch across cleavage fractures and parallel calcite veins (photo 76). Both dextral and sinistral offsets occur accompanied by a smaller vertical component. The displacements are observed by noting the traces of intersecting calcite veins across the slip planes. A count of the observed sense of slip indicated that 25 out of 35, i.e. approximately 70% of such slips, are sinistral.

Calcite veins also fill dip joint fractures in rocks.

Rare strike sets of calcite veins roughly perpendicular to the bedding cross the Ordovician rocks. Such sets displaced across bedding in concentric  $F_2$  folds suggest that they formed before full development of the folds (see page 37 and photo 94).

Some calcite veins are slightly oblique to the  $S_2$  cleavage and cross with no offsets or deformation from one plane to the next (photo 71), but on average the two are coincident. Their general parallelism and undeformed nature indicates that they post-date the anisotropy along the  $S_2$  cleavage.

It is considered that the development of the axial plane calcite veins is somewhat analogous to the calcite enrichment along the  $F_2$  axial planes (see below). It post-dated an early stage in the formation of the cleavage but was formed before the  $F_2$  folds were fully formed (see section, Slickensides on Calcite Veins).

Calcite Concentration Parallel to the  $F_2$  Axial Planes. Secondary concentrations of calcite often occur parallel to the  $F_2$  axial planes. These are linear and irregular patches in the hinge areas of the  $F_2$  folds in calcareous slates of Cambrian and Ordovician age (photo 26).

In some examples the calcite concentrations are seen under the microscope to be slightly discordant to  $S_2$  cleavage fractures (microphoto 18). The original sedimentary layers are marked by a difference in crystallinity in the calcite ground mass.

In spite of local discordance with the cleavage on a microscopic scale, the calcite rich zones lie approximately parallel to the axial plane and  $S_2$  cleavage on a larger scale.

The concentrations are considered as formed by enrichment and replacement of the rock material by calcite along fold hinges. The general parallelism with the axial planes may indicate that the calcite migrated at the end of the folding, when, due to residual buckling stresses, the hinge area would be a 'low' pressure area.

Slickensides on Calcite Veins.

The axial plane calcite veins are extensively slickensided and contain two dominant sets: One plunges parallel to the southeasterly dip direction of the calcite vein (with a pitch of  $90^\circ$ , photo 72), the other southwesterly or northeasterly at low angles. A few slickensides plunge diagonally across the veins, and in still fewer cases slickensides with multiple directions were observed (photo 75).

The age relations of the two major sets are ambiguous because each film of calcite in the vein contains only one set of slickensides. Obliteration of one set by the other could not be used to determine the relative age of the two. The predominance of the steeper set over the low plunging variety probably indicates that the major slip on the veins was parallel to the dip direction.

In plate V the slickenside pitches as measured on the calcite veins are plotted on a Wulff net. This net, rather than the equal-area net, was used as the diagrams were not contoured, and plunge variations in low plunging linear features are more readily apparent. The plunges of the axes of the  $F_2$  folds in each locality were also plotted on the same net.

In the diagrams the slickensides and the fold axes tend to lie on the same plane. Since the variably plunging fold axes define the local axial plane, the slickensides also tend to lie on the average  $S_2$  plane.

The diagrams (plate V) indicate that the slickensides are independent of the plunge of the  $F_2$  folds. In many localities, whereas the  $F_2$  fold plunge is variable, the dominant slickenside cluster pitches at approximately  $90^\circ$  on the  $S_2$  surface. In some localities slickensides with steep to low northeasterly pitches are also common (Plate V; ME 11, 11a; MW 7).

It is considered that the steep slickensides on the calcite veins are the result of major movement on the vein surfaces during  $F_2$  folding. They are related to the development or accentuation of slip on the  $S_2$  cleavage planes.

The subhorizontal slickensides are more difficult to explain, but are relatively few and can be related to the sinistral offsets observed across calcite-filled cleavage planes (photo 76). The following factors are listed in favour of this view:

(i) The sinistral displacements across the cleavage and parallel calcite veins, and presumed lateral slip giving rise to low pitching slickensides on the calcite veins can be ascribed to the same sense of slip. The sinistral slip is probably associated with predominantly S-shaped  $F_3$  folds (p. 56).

(ii) The relatively few examples of low pitching slickensides correspond well with the few strike-slip displacements observed in the area. Diagonal multiple sets of slickensides are regarded as a result of diagonal movement between dip and strike.

A set of slickensides on the calcite veins and occasionally on bedding surface of the slate underlying the siltstone or limestone layers, parallels the dip of the beds. These are restricted to the competent and non-competent interbeds, and are considered as due to concentric bedding slip accompanying  $F_2$  folding.

### Bedding Foliation

Some rocks especially those of Ordovician age along the St. Lawrence coast contain a prominent parting parallel to the bedding. The rock type with this foliation is usually shale with fine laminations of limestone or siltstone or minor shale enclosed in competent beds. This is a sedimentary feature.

However, a similar bedding foliation also develops within argillaceous rocks in the proximity of massive or competent beds. Along the coast of the St. Lawrence River at localities near Les Mechins it is well demonstrated mesoscopically in finely interbedded shale, limestone, and siltstone, and near Ste. Felicite in shale interbedded with impure grey sandstone. Microscopically the foliation is apparent in interbeds of shale at contacts with siltstone layers where micas are aligned parallel to bedding (microphoto 5 and 6 contain this feature, but do not show it clearly). Mica alignment is strongest in tight folds (photo 97).

The intersection of bedding foliation with the axial plane cleavage  $S_2$  could indicate that the foliation developed earlier than the cleavage, perhaps due to earlier folding as for the Shickshock rocks. However, the restriction of the foliation to the parts of the argillaceous layers next to more competent beds suggests that it is the result of slip and shearing at the contacts during  $F_2$  folding, and thus is penecontemporaneous with  $S_2$ .

### Regional Variation in Style of $F_2$ Structures

$S_2$  Cleavage in Different Rock Formations. The  $S_2$  cleavage development in different rock types is given in table 5. The table shows that the

Table 5. TABLE SHOWING NATURE OF AXIAL PLANE FRACTURE CLEAVAGE  $S_2$  IN DIFFERENT ROCK FORMATIONS:

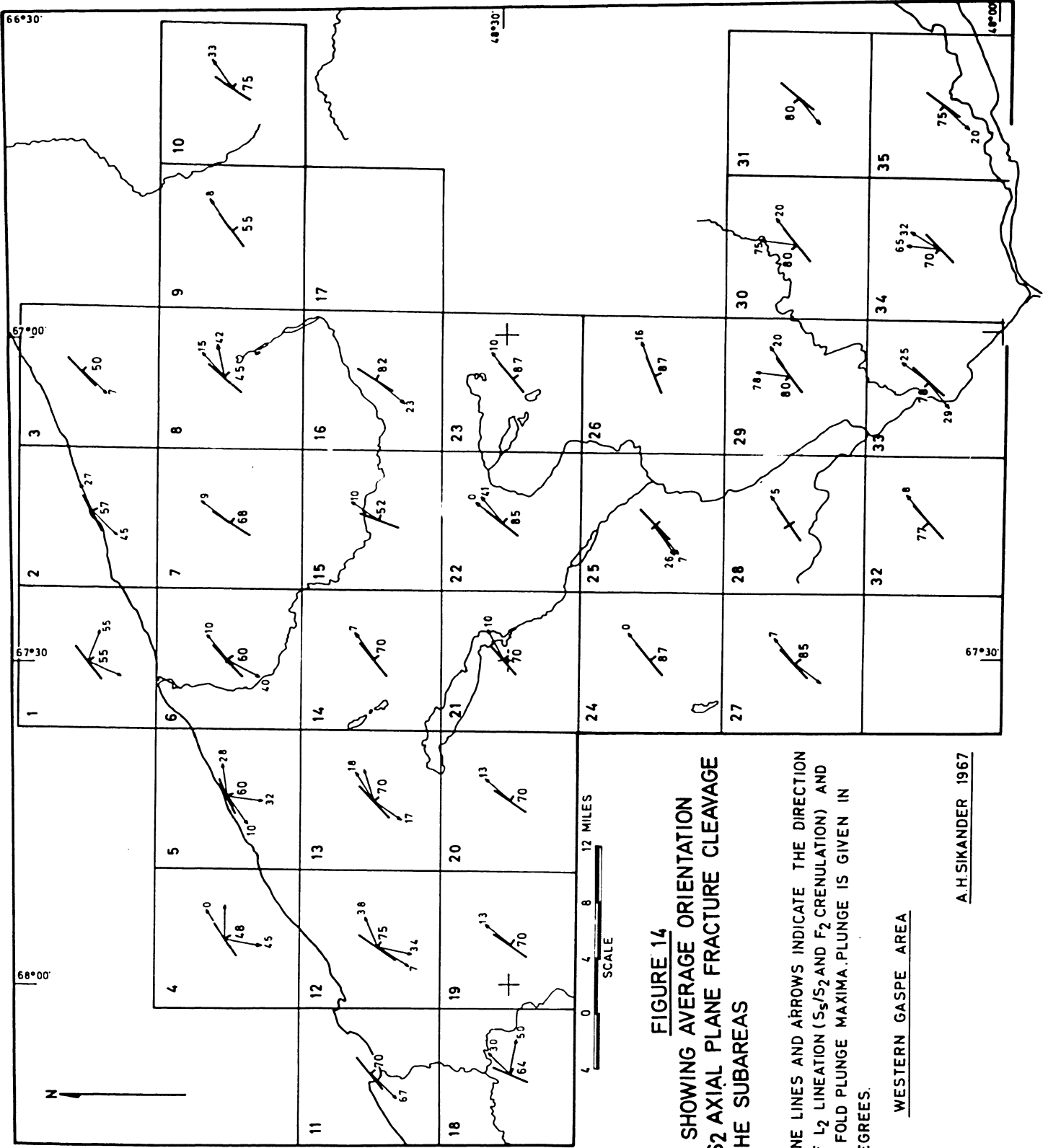
Rock Formations Groups	Degree of Development	Smallest Penetrative Domain	d/D Approx. Average	Remarks
YORK RIVER HEPPEL	Poor, Coarse	Macroscopic	<.25	
FORTIN SLATE	Good, Variable	Mesosopic Microscopic	<.25 >.25	
GRANDE GREVE	None- Poor Coarse	Macroscopic	<.25	
CAPE BON AMI	Poor - Coarse	Macroscopic	<.25	
ST. LEON	Poor to Fair in Shale	Macroscopic Microscopic	<.25	
SYABEC	Poor	Macroscopic	<.25	
VAL BRILLANT	None to very Poor	Macroscopic	<.25	
AWANTJISH	None	-	-	
INCOGNITO CONGL.	None	-	-	
QUEBEC	Good - Poor	(Microscopic (Mesoscopic (Macroscopic	<.25 >.25	Crenulation of micro- lithons
KAMOURASKA FACIES	Poor - None	Macroscopic	<.25	
SHICKSHOCK	Poor - None	Microscopic	-	
CAMBRIAN- ORDOVICIAN	Good - Poor	Microscopic	<.25 >.25	Crenulation of micro- lithons

d = Displacement across cleavage surface.

D = Spacing between cleavage surface.

Poor = Impersistently developed as observed in field.

Good = Well developed as observed in field.

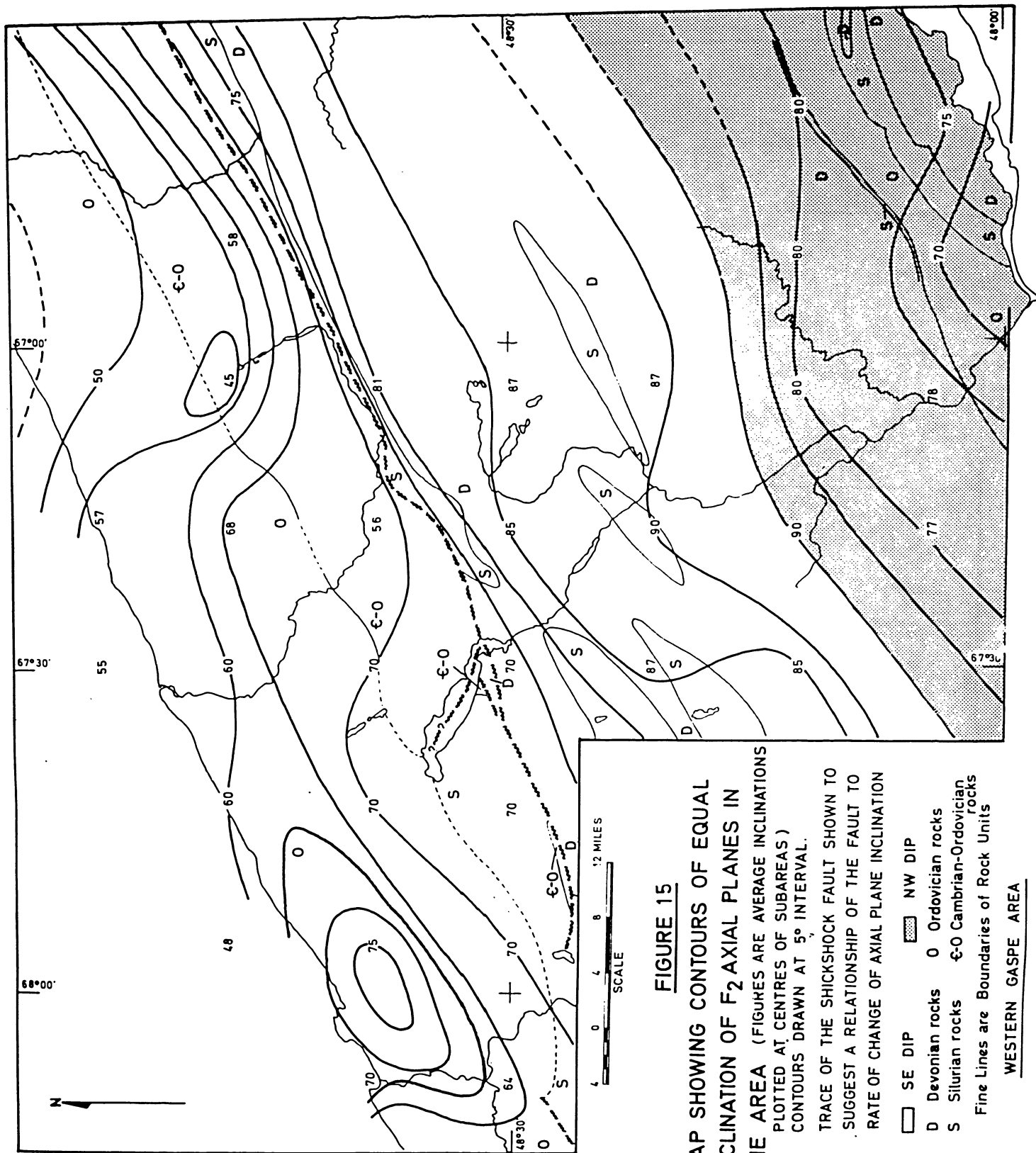


**FIGURE 14**  
**MAP SHOWING AVERAGE ORIENTATION**  
**OF S2 AXIAL PLANE FRACTURE CLEAVAGE**  
**IN THE SUBAREAS**

FINE LINES AND ARROWS INDICATE THE DIRECTION OF L<sub>2</sub> LINEATION (S<sub>1</sub>/S<sub>2</sub> AND F<sub>2</sub> CRENULATION) AND F<sub>2</sub> FOLD PLUNGE MAXIMA. PLUNGE IS GIVEN IN DEGREES.

WESTERN GASPE AREA

A.H. SIKANDER 1967



**FIGURE 15**  
**MAP SHOWING CONTOURS OF EQUAL**  
**INCLINATION OF F<sub>2</sub> AXIAL PLANES IN**  
**THE AREA (FIGURES ARE AVERAGE INCLINATIONS**  
**PLOTTED AT CENTRES OF SUBAREAS)**  
**CONTOURS DRAWN AT 5° INTERVAL.**

TRACE OF THE SHICKSHOCK FAULT SHOWN TO  
 SUGGEST A RELATIONSHIP OF THE FAULT TO  
 RATE OF CHANGE OF AXIAL PLANE INCLINATION

- SE DIP    ▤ NW DIP
  - D Devonian rocks    O Ordovician rocks
  - S Silurian rocks    ε-O Cambrian-Ordovician rocks
- Fine Lines are Boundaries of Rock Units

WESTERN GASPE AREA

axial plane cleavage is well developed in finer lithologies such as the Ordovician slates of the Quebec Group and the Devonian Fortin Slates. The former are exposed along the northern coastal area, and the latter in a belt along the Gaspé synclinorium and south of it. In these rock types the cleavage is penetrative on mesoscopic or microscopic scales and develops as slip or fracture cleavage. Curvature in fine layers in cleavage fracture zones seen under the highest magnification in an ordinary microscope may be due to submicroscopic displacements. In the coarse rock types the cleavage is poorly developed, is penetrative on a macroscopic scale and is of a fracture variety.

Although the  $S_2$  cleavage in the area is present in all rock types irrespective of age other than very coarse clastics, it is virtually absent in rocks of the Shickshock Group. These rocks contain a strong early schistosity  $S_1$  which is deformed by  $F_2$  folds but rarely fractured parallel to the  $F_2$  axial planes.

Regional Variation in Attitude of Axial Planes and  $S_2$  Cleavage. It has been established by the study of the  $S_2$  cleavage and the fold elements that the cleavage is parallel to the axial planes of the  $F_2$  folds (plate IVa and see Axial Plane Cleavage). Mean attitudes of the  $S_2$  fracture, therefore, corresponds to the attitudes of the axial planes of the  $F_2$  folds in the subareas.

The map in figure 14 shows the average inclinations of the  $S_2$  cleavage in each subarea. The map was obtained by plotting the poles to the  $S_2$  cleavage on polar stereonet for each subarea. The centre of gravity of each polar maximum was plotted at the centre of each subarea. The map represents a total of 2908 observations on cleavage surfaces.

Figure 15 (contoured from figure 14) shows that in the northern area the cleavage dips towards the southeast and the lowest mean inclinations are around  $50-55^\circ$ .  $S_2$  gradually attains a vertical position along a central belt. Further south it dips northwest, where the lowest mean inclination is  $70^\circ$ . A section across the general strike of the area demonstrates the upward divergence of the  $S_2$  cleavage (figure 6b). This is analogous to the upward fanning of  $S_2$  in the

synclines in the mesoscopic and macroscopic scales. The upward fanning is referred to as Gaspé Cleavage Fan, across the Gaspé Synclinorium.

North of the central belt of vertical cleavage and upright axial planes there is a belt mainly consisting of Cambrian and Ordovician rocks, where there is a rapid change of axial-plane inclination. The change further north is more gradual. The belt of rapid change is in the vicinity of the Shickshock Fault. Near Matapédia Lake, where the fault splits into smaller faults, probably with a considerable decrease in the throw, the contour lines disperse, showing a decreased rate of change (see the Shickshock Fault).

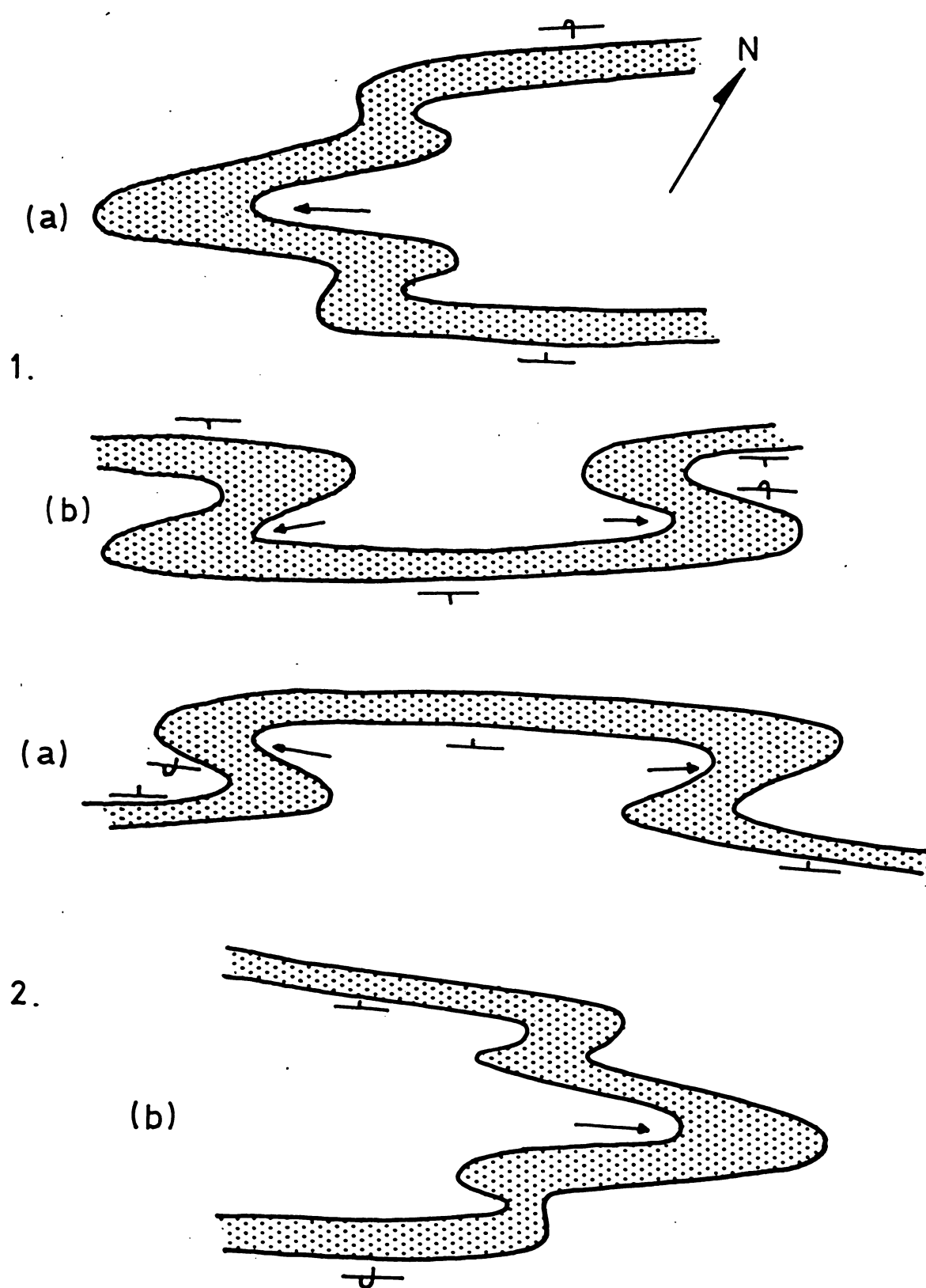
Since the axial plane inclination from each 100 sq. mi. subarea are plotted at the centre of these areas (fig. 15), the mean attitudes plotted immediately south of fault include measurements from folds that occur north of it. Therefore, although the contours in fig. 15 are relatively close together south of the fault, suggesting a rapid areal change in inclination, the field observations indicate that most of the change is restricted to the north of the fault.

Division of the Area into Major Domains with Respect to  $F_2$  Fold

Profiles and Attitudes. The area under investigation has been divided into three major structural domains with respect to the  $F_2$  folds (table 6).

Table 6. Division of the Area into Domains according to Fold Limb Ratio and Axial Plane Attitude.

Domain	Limb Ratio of Anticlines	Attitudes of Axial Planes	Rock Type	Geographical Distribution
I.	Shorter limb to NW. Up to 1:3	50°SE to subvertical	Cambrian and Ordovician	Area north of Shickshock Fault
II.	Shorter limb to NW. Approx. 1:1 to 1:2	Vertical or Subvertical	Silurian Devonian	Area lying between the Shickshock Fault and axial area of synclinorium.
III.	Shorter limb to SE. 1:1 to 1:2	60° NW to Subvertical	Upper Ordovician, Silurian & Devonian	Area south of axial area of Synclinorium



**FIGURE 16.** S AND Z-SHAPED OUTCROP PATTERNS OF OVERTURNED PLUNGING  $F_2$  FOLDS. 1, OVERTURNED NORTHWARD (a) UNIFORMLY PLUNGING PARASITIC FOLDS ON OPPOSITE LIMBS OF A PLUNGING FOLD (b) DOUBLY PLUNGING FOLDS. 2, OVERTURNED SOUTHWARD (a) DOUBLY PLUNGING FOLDS (b) UNIFORMLY PLUNGING PARASITIC FOLDS ON OPPOSITE LIMBS OF A PLUNGING FOLD.

It is suggested that domains I and II belong to a larger structural domain, both forming part of the northwestern limb of the Gaspé Synclinorium (fig. 6a). The difference between I and II lies in the degree of overturning of the axial planes of the  $F_2$  folds.

Outcrop Pattern in the Domains. Double plunging, asymmetrical, overturned  $F_2$  folds of unequal limb lengths give rise to S or Z shapes on horizontal sections depending on whether the fold is plunging northeasterly or southwesterly and on the direction of overturning. In domains I and II, where folds are overturned northward, the southwesterly plunging end of a double plunging fold makes a Z shape in an outcrop pattern and a S shaped pattern is made by the northeasterly plunging end of the folds (fig. 16, 1b.). In contrast, in domain III, where folds are overturned southward, northeasterly plunging folds give rise to a Z shaped plan and southwesterly folds make an S shaped plan (figure 16 2.a).

Parasitic folds on opposite limbs of a larger fold give rise to varying outcrop patterns (fig. 16 1.a, 2b). The parasitic folds plunging in the same direction give rise to identical shapes in domains I and III.

S shaped patterns are also formed by regional fold structures, as can be seen in the geological map (fig. 6 south of Casault Lake). The varying shaped outcrop patterns in the Lower Paleozoic rocks are, therefore, due to a single phase of deformation ( $F_2$ ).

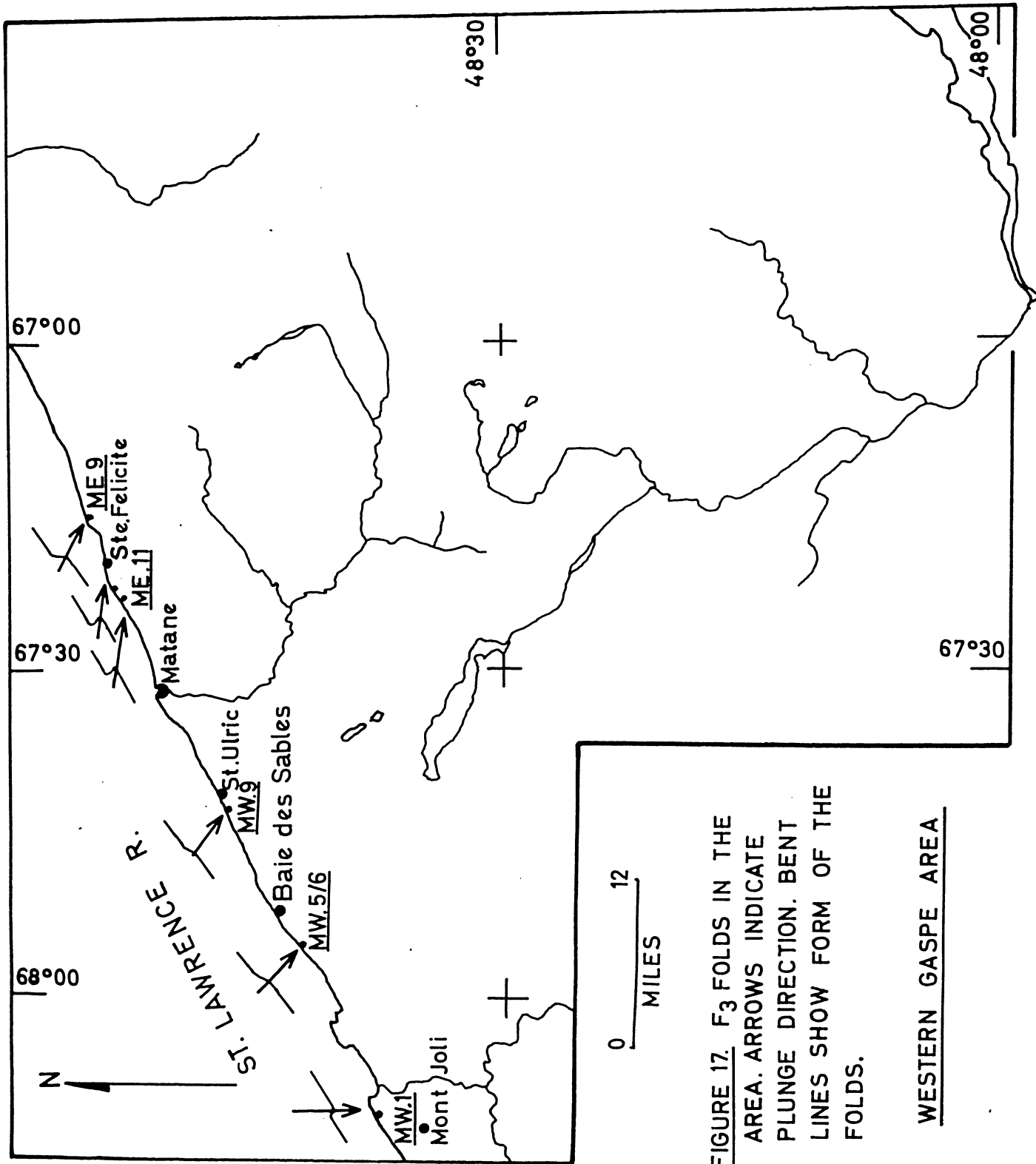
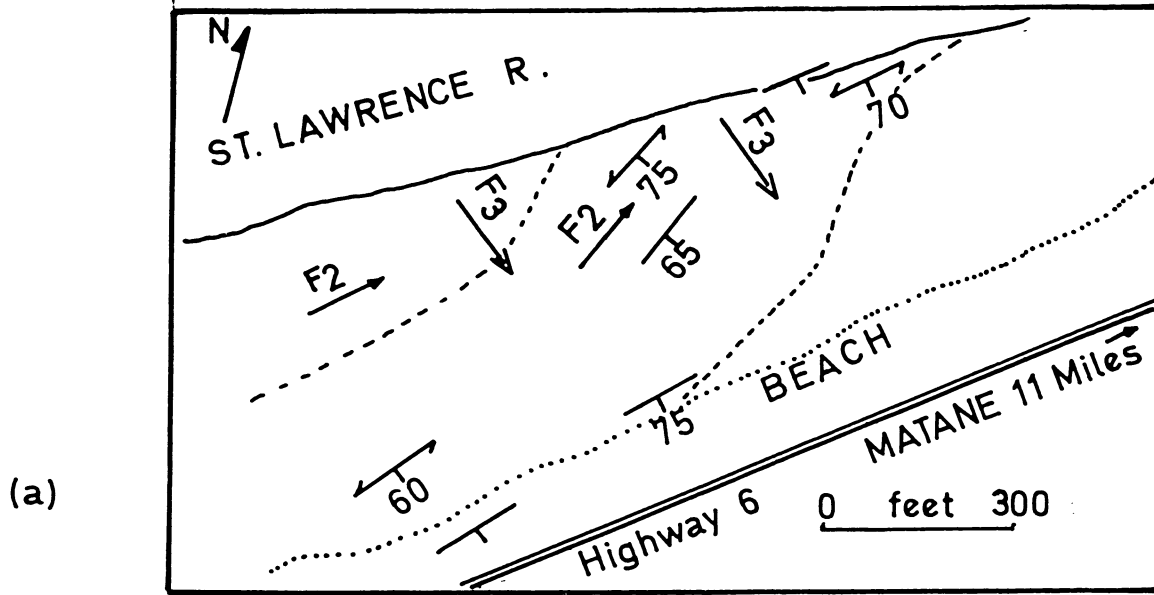
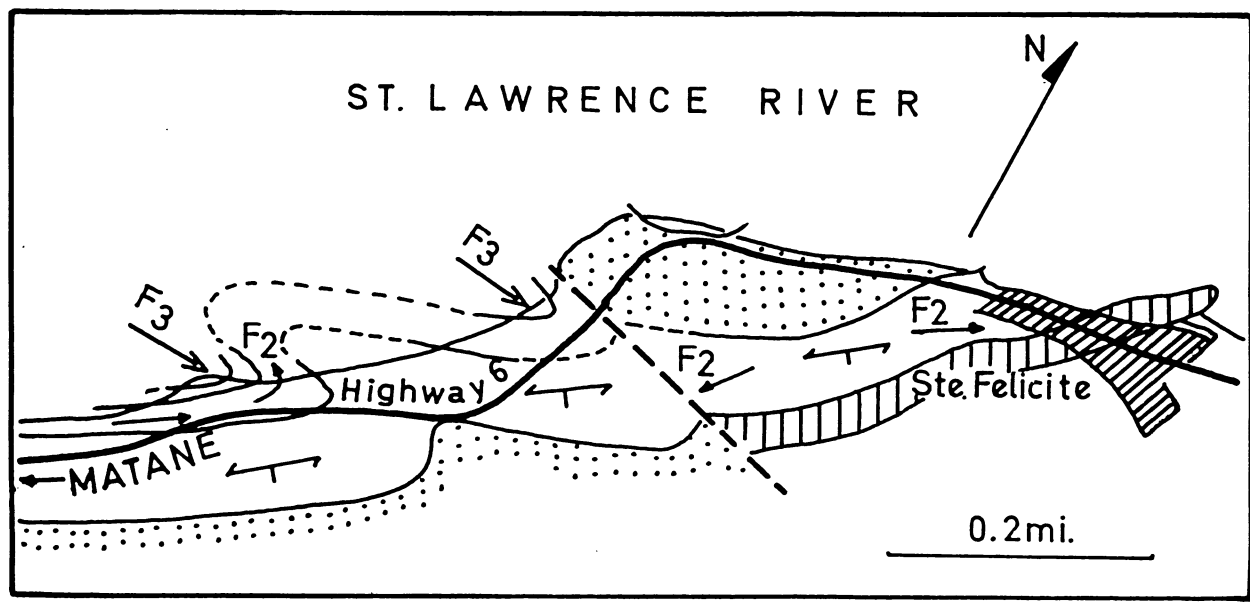


FIGURE 17. F<sub>3</sub> FOLDS IN THE WESTERN GASPE AREA. ARROWS INDICATE PLUNGE DIRECTION. BENT LINES SHOW FORM OF THE FOLDS.

WESTERN GASPE AREA



(a)



(b)



Figure 18. (a) Open F<sub>3</sub> folds exposed 11 miles SW of Matane, MW-9, plunge steeply SE.  
 (b) Acute F<sub>3</sub> folds exposed SW of Ste Félicité. Outcrop pattern interpreted from low tide exposure and aerial photographs.

### 3. The F<sub>3</sub> Folds.

#### Introduction and Distribution

F<sub>3</sub> cross folds deform the F<sub>2</sub> structures in the northern coastal sections. They were not observed in the interior. These folds occur irregularly on a macroscopic scale, and a few are mesoscopic.

F<sub>3</sub> folds crossing F<sub>2</sub> are best exposed one mile southwest of Ste. Felicite village, ME-11 (fig. 17). The folds are exposed at low tide and can be followed in aerial photographs. In this locality S<sub>2</sub> is bent by F<sub>3</sub> folds (fig. 18b). Other localities where F<sub>3</sub> are exposed are: Four miles NE of Ste. Felicite (ME-9), an F<sub>2</sub> anticline in the impure grey sandstone is overturned to the northwest and re-folded by F<sub>3</sub> in a S-shaped pattern. Eleven miles southwest of Matane, southwest of St. Ulric (photo 78 and field sketch figure 18a) and also along the coast northeast and southwest of Baie des Sables 18½ miles SW of Matane (MW-5/6).

#### Profile

The form of the F<sub>3</sub> folds varies from a broad S-shaped flexure as observed southwest of St. Ulric (photo 78) to acute S-shaped forms as near Ste. Felicite (fig. 18b). Typically the folds are concentric and rounded in profile. A Z-shaped, probably F<sub>3</sub> cross fold was observed along the coast of north of Mont Joli (MW-1, photo 90).

The F<sub>3</sub> folds deform the S<sub>2</sub> cleavage, and cause a scatter of the poles to the S<sub>2</sub> planes in a stereonet plot (diagram MW-9, Plate IVa shows a scatter in S<sub>2</sub> cleavage and axial plane poles which is attributed to F<sub>3</sub> folding).

#### Planar and Linear Elements

The axial surfaces of the F<sub>3</sub> folds are planar and vertical or subvertical and strike southeast or east. A coarse and irregular fracture which resembles a closely spaced joint set is occasionally

present in the hinge area of the folds.

The plunge of the  $F_3$  folds, determined by the southeast dips of  $S_1$  and  $S_2$  is to the east or southeast. An axial lineation is expressed by the intersection of  $S_2$  and  $S_3$ , but this lineation is rare.

The shape and orientation of the acute 'S' folds suggests that they may have formed in conjunction with sinistral slip acting in a SWS-ENE direction (see p. 49 and fig. 26).

Sinistral strike slip movement giving rise to subhorizontal slickensides on the calcite veins in the coastal area (calcite veins p. 46) area could have accompanied the  $F_3$  folding. Dip slickensides on the same calcite veins are considered as due to the  $F_2$  folding. Possibly the horizontal slip, slickensiding, and folding closely followed the  $F_2$  deformation.

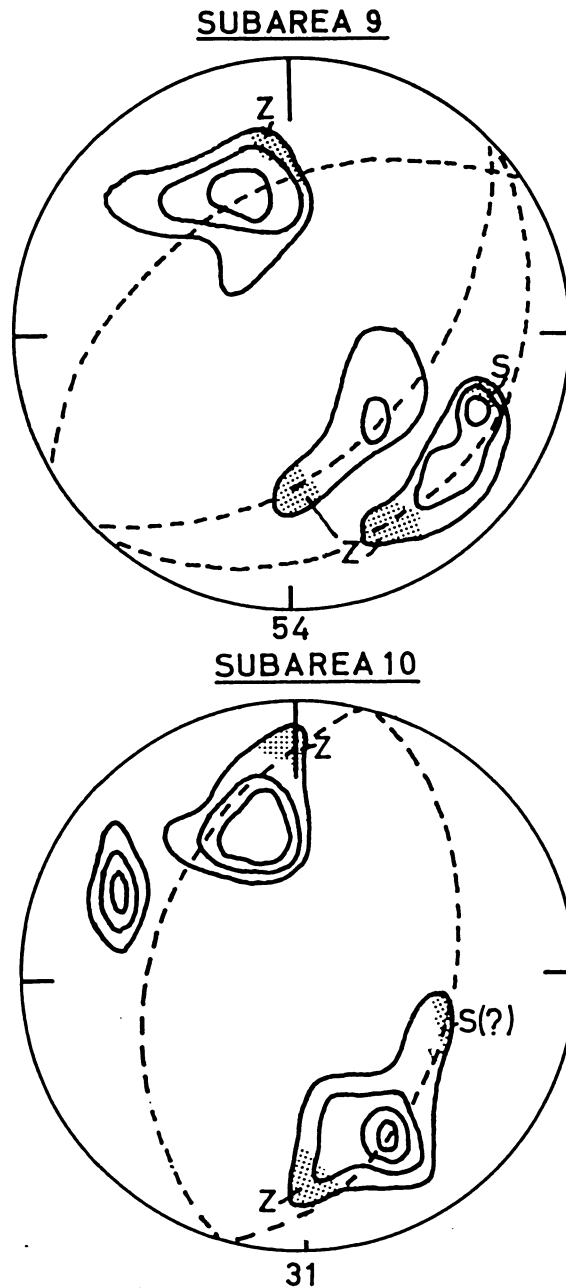


Figure 19. S and Z-shapes of  $F_4$  folds in relation to pitch on  $S_1$  surface. Mean  $S_1$  surfaces on which  $F_4$  are measured are shown as dashed great circles. Folds plunging more N-S are Z-shaped and those plunging more E-W are S-shaped. Shaded areas indicate the predominant shapes observed. Few S-shaped folds were noted plunging ESE in subarea 10. The  $F_4$  plots on equal area net are contoured at 2, 4, 6, and 10 per cent frequencies per 1 per cent area.

#### 4. The $F_4$ Folds

##### Introduction and Distribution

$F_4$  folds, mesoscopic and microscopic in scale deform  $S_1$  surfaces which are folded by  $F_2$  folds, and resemble kink bands. They are usually restricted to the limb areas of the  $F_2$  folds. The folds are regular in orientation on outcrop scale but on a larger scale they are variable.

Most  $F_4$  folds occur in the Shickshock Mountain area where  $S_1$  foliation is present (fig. 9). They are common in metavolcanic, less frequent in associated metasedimentary rock, and rare in the belt of Cambrian-Ordovician phyllitic slates. MI-1/5 (photo 19) and AMQ-10/3 (photo 20) are the two localities (fig. 2) in grey phyllitic slates to the north of the Shickshock Mountains in which  $F_4$  axes were identified. Matapedia volcanic rocks correlated with the metavolcanics in the Shickshock Mountains (Ollerenshaw 1963), do not contain the  $F_4$  structures. A lack of well defined schistosity in the metavolcanics and metasediments near Matapedia Lake probably accounts for the absence.

$F_4$  folds are absent from the grey slates interbedded with metavolcanic and metasedimentary rocks that contain  $F_1$  folds (localities in intertonguing slates 3 miles east of Depot Jean fig. 6).

The  $F_4$  folds are not observed in the Ordovician and younger rocks in the area.

##### Profile

The  $F_4$  folds are open, angular and symmetrical to asymmetrical in shape. The asymmetrical folds occur in S or Z-shapes as viewed down the plunge (photo 17, Z-shape; and photo 19, S-shape; and microphoto 19, S-shape).

In many examples the direction of pitch and trend of the axes on the folded schistosity surface  $S_1$  determines the S and Z shape of the folds. The S-shaped folds have a southeasterly trend

and the Z-shaped folds a north-south trend (fig. 19). The two shapes of the folds do not occur together in one locality. The folds pitching down the dip are generally symmetrical with shorter wave length.

Figure 19 illustrates only the  $S_1$  plots that contain the  $F_4$  folds. Hence the northerly trending attitude of  $S_1$  in figure 19, subarea 10, is not the representative attitude of  $S_1$  in that subarea (see plate I).

The wave length scale is variable geographically. Southwest of Matane Lake the fold wave length varies between .1 and .3 inches. In the Cap Chat river area to the northeast separation is irregular, .1 to .5 inch apart in the south, and less in the northern margin of the Shickshock Mountains. East of the study area, near South Mountain, the  $F_4$  folds are symmetrical and regularly spaced. Further east near Mount Albert near La Gite, hotel of the Gaspé National Park, the  $F_4$  were not observed in the metovolcanics, but in the gorge of Ste. Anne River they are consistently developed.

#### Planar and Linear Elements

The linear expression of the folds consists of small corrugations on  $S_1$  surfaces (photos 15-17) which are best observed in oblique illumination. These form at large angles to the  $F_1$  and  $F_2$  fold axes marked by  $L_1$  and  $L_2$  lineations.  $L_4$  is defined by the intersection of  $S_1$  schistosity with axial planes of the  $F_4$  folds.

The  $F_4$  folds pitch down both the limbs of the  $F_2$  folds. The southeasterly plunge in the nets (plate VI) is dominant over the northwesterly one in view of the predominance of the outcrops on the longer southeasterly anticlinal limbs of  $F_2$  folds.

Owing to the small scale of the  $F_4$  folds it was impossible to measure many axial surfaces. Field observations where  $L_4$  varies across  $S_1$  folded by  $F_2$  indicate that the axial surfaces containing  $L_4$  are planar and subvertical. Although they are not observed together on a single outcrop the S and Z shaped folds can be regarded as pairs of conjugate folds (Paterson and Weiss 1966), formed by a NE-SW compression parallel to the foliation (Ramsay 1962a, fig. 27, see p. 81).

Age of the F<sub>4</sub> Folds

The age of the F<sub>4</sub> folding is deduced from the following:

(i) The F<sub>4</sub> folds deform the S<sub>1</sub> schistosity, therefore they are not post F<sub>1</sub> folding.

(ii) The absence of F<sub>4</sub> folds from the Ordovician and younger fine grained and well cleaved rocks may possibly suggest that F<sub>4</sub> pre-dates these rocks. However this can be ascribed to geographic distribution or lack of a fine anisotropy (S<sub>1</sub> foliation) in the rocks.

(iii) Axial surfaces of F<sub>4</sub> appear planar across S<sub>1</sub> deformed by F<sub>2</sub>.

The F<sub>4</sub> folding is therefore considered as post F<sub>2</sub>.

The relative age between the F<sub>3</sub> folds and F<sub>4</sub> is not clear.

However, the following is relevant:

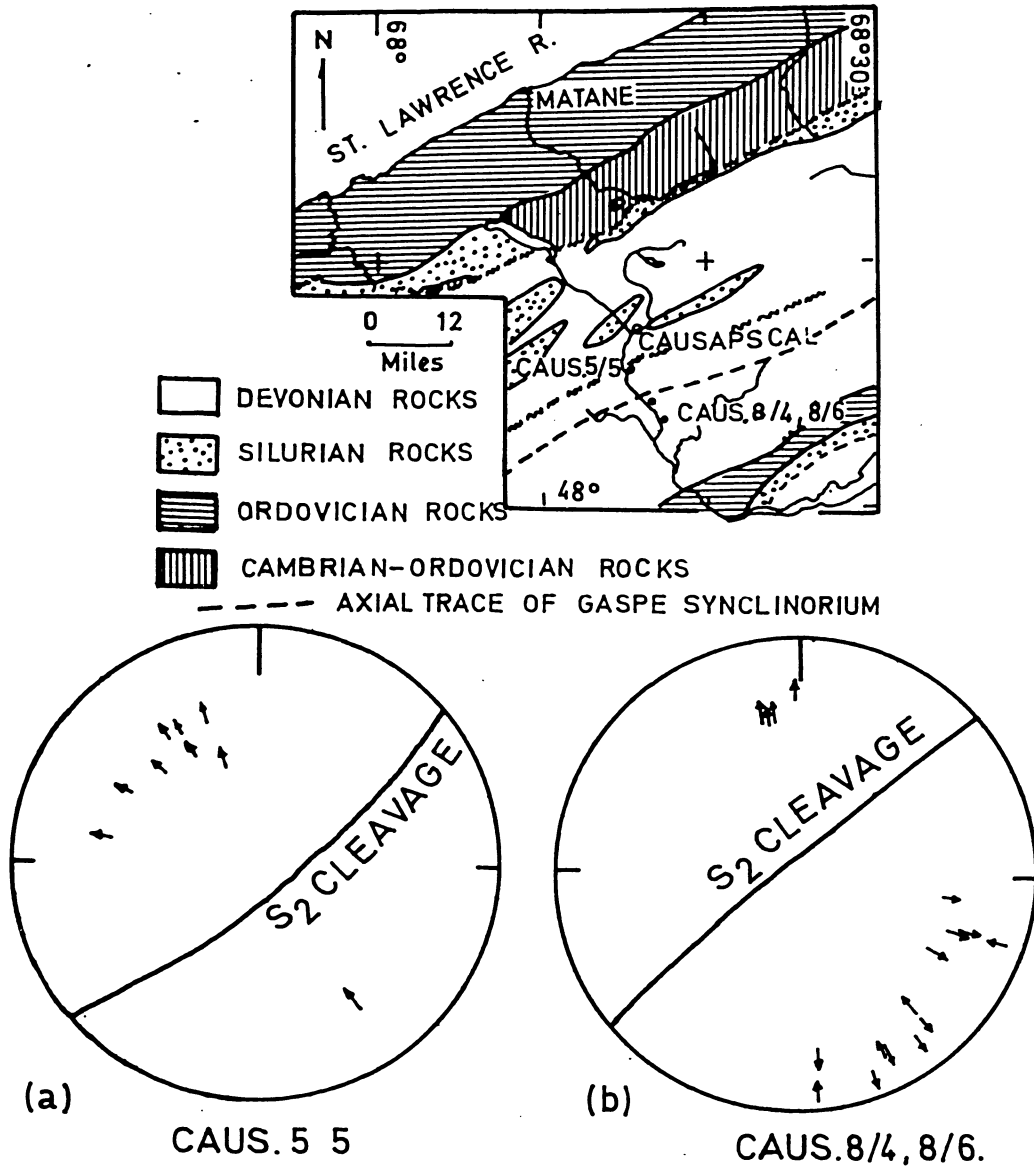
(i) The F<sub>3</sub> folds are restricted to the northern part of the area, and the F<sub>4</sub> folds to the Shickshock Mountain area.

(ii) There is a suggestion that F<sub>3</sub> closely followed the F<sub>2</sub> folding (page 57).

(iii) Whereas F<sub>3</sub> are S-shaped, the F<sub>4</sub> folds are both S and Z-shaped. This suggests a difference in the mechanism of formation and therefore in age. The F<sub>3</sub> folds are mesoscopic and macroscopic and F<sub>4</sub> are mesoscopic and microscopic. The difference in scale could be due to differences in anisotropy spacing (F<sub>3</sub> does not affect rocks with S<sub>1</sub> foliation), rather than a change in mechanism.

It is suggested that the F<sub>4</sub> folding post-dated the F<sub>3</sub> folding.

However, it is possible that F<sub>3</sub> and F<sub>4</sub> folds were pene-contemporaneous.



**Figure 20.** Location and orientation of late stage kink bands. Position of the arrowheads indicate dip of the kink planes on Wulff net. Arrows indicate the displacement of the upper part of the kinks. The two diagram localities are as shown in map. Average  $S_2$  cleavage is shown in the diagrams.

## 5. Late Stage (?) Kinks and Associated Structures

The following structures are included as late stage structures:

- (i) Low dipping kink bands.
- (ii) Low dipping slip planes and micro-kinks.

### Low Dipping Kink Bands

Kink bands occur in well cleaved Lower Paleozoic rocks. They are most frequent across steep  $S_2$  cleavage in the non-calcareous slate formations of the Devonian Fortin Group, and are well exposed in a few localities along the central part of the area in the Matapedia Valley (fig. 20, photo 88).

The kink bands vary in width from an inch or less to six inches, and the average length along the axis is approximately five feet. The axes are subhorizontal, parallel to the strike of the  $S_2$  cleavage. The kink planes are low dipping (between  $5-36^\circ$ ) to the northwest or southeast (fig. 20). The upper parts of the kink bands are displaced horizontally relative to the lower parts, as indicated on the Wulff nets in figure 20. Since most of the planes were low dipping and no contouring was necessary, the Wulff net was used to avoid getting the diagrams crowded towards the periphery.

On the northern limb of the Gaspé synclinorium where  $S_2$  cleavage dips steeply southeast, nearly all the kink planes dip gently northeast and the upper parts are consistently displaced in this direction (fig. 20a).

To the south, in localities Caus 8/4 and 8/6, the average dip of the  $S_2$  cleavage is  $85^\circ$  northwestward (fig. 20b). The two localities are thus situated within a belt of subvertical to vertical axial planes. Therefore, although the localities on fig. 20 appear south of a clearly defined axial trace of the synclinorium, they are, in view of the average inclination of the local folds, considered as situated nearly within the axial area. In these localities most kinks dip southeastward, and both northwesterly and southeasterly displacements occur. In a few localities smaller (narrower) kinks lie within larger ones (photo 88). The sense of displacement in both is the same, but the smaller ones form at an angle of  $15-20^\circ$

to the larger, and the acute angle points in the direction of the displacement of the upper parts.

It is considered that the displacements across the low dipping kink bands are related to late stage deformation of the Gaspé synclinorium. On the northern limb movement of upper rocks was directed away from the axial area, and immediately to the south of the axis the movement direction was variable.

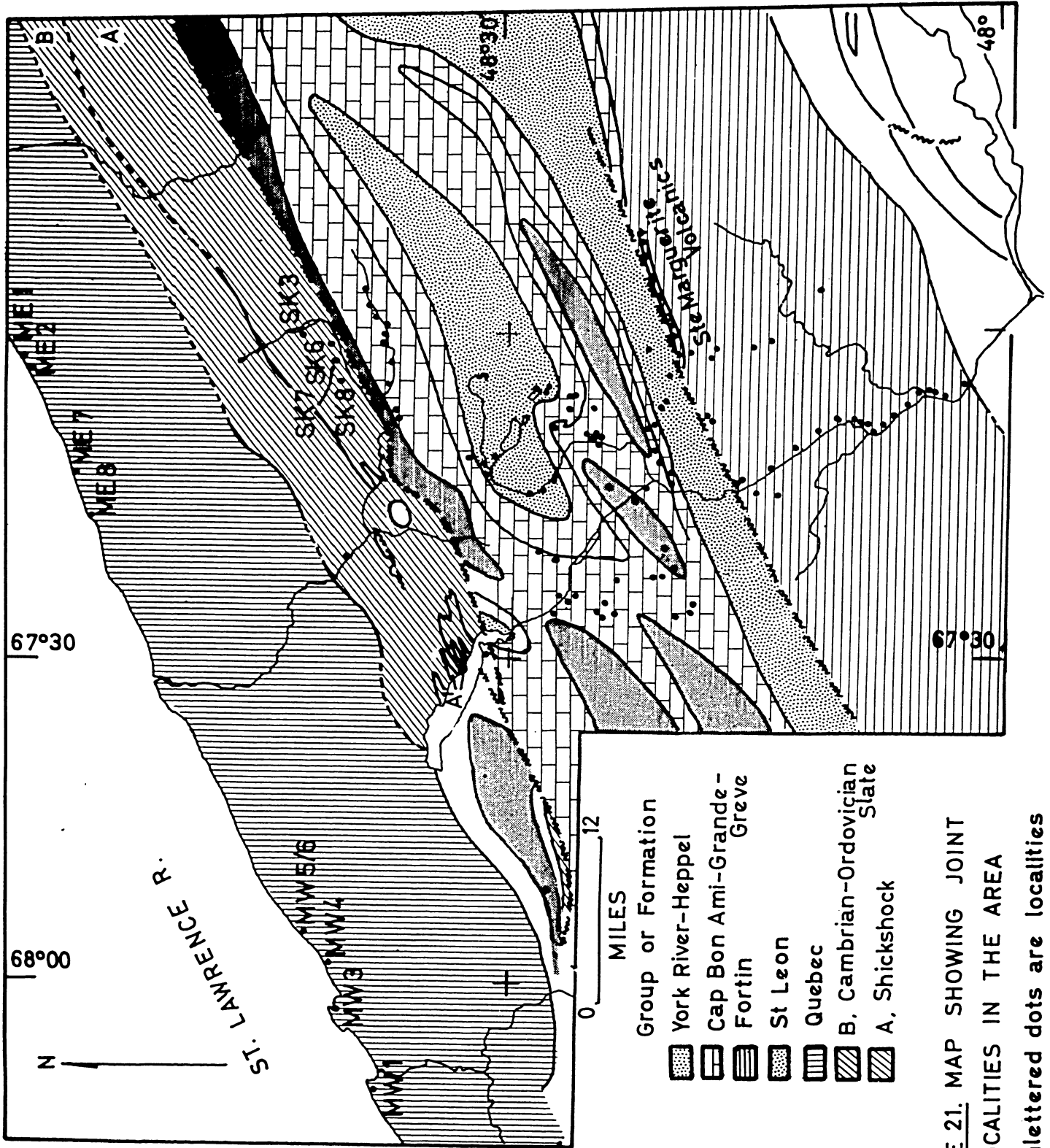
#### Low Dipping Slip Planes and Micro-Kinks

These occur in well cleaved and fine grained Cambrian, Ordovician and Devonian rocks on the northern limb of the Gaspé synclinorium (domains I and II, page 54). The Silurian rocks, being in general coarse grained and competent, lack these structures. The slip planes consist of irregularly spaced non-penetrative weak fractures which form at large angles to steep  $S_2$  cleavage. The fractures do not cut across contrasting lithologies. They are commonly characterised by microscopic displacements with upper part usually shifted to the northwest.

The fractures dip at low angles to the northwest and strike northeast or acutely oblique to the strike to the  $S_2$  cleavage.

A lineation  $L_L$  is defined by the trace of the intersection of  $S_L$  and  $S_2$  cleavage. A similarly oriented lineation is formed by irregularly spaced microscopic angular monoclinial folds. Usually these lineations are restricted to a part of an outcrop. In the Ordovician rocks  $L_L$  and  $S_L$  are best observed in road sections between Portage Lake and Towago Lake (1 mile southwest of Portage Lake) and up to 2 miles southwest of Portage Lake (fig. 2 and 6). They also occur in Cambrian-Ordovician phyllitic slates 8 miles west, and 9 miles southwest of Depot Jean, and in Fortin slates along the Matapedia River 1 mile south of Ste. Florence Fault, where they are within 300 feet, but not adjacent, to kink bands (fig. 20a).

The slip planes and micro-kinks are regarded as contemporaneous with the kink bands because they are approximately parallel, and as observed in the field, they displace the upper rocks in the same direction.



**FIGURE 21. MAP SHOWING JOINT LOCALITIES IN THE AREA**

Unlettered dots are localities in Siluro-Devonian rocks.

## 6. Joints

### Introduction

A study of joints was made in all rock types in the Lower Paleozoic succession. Each locality investigated (fig. 21) was used for data on the study of local joint patterns.

The observations for each rock type were plotted separately to study the influence of lithology and stratigraphic position on the joint pattern (plates VII). An attempt was also made to compare the joint development with the degree of folding and cleavage development.

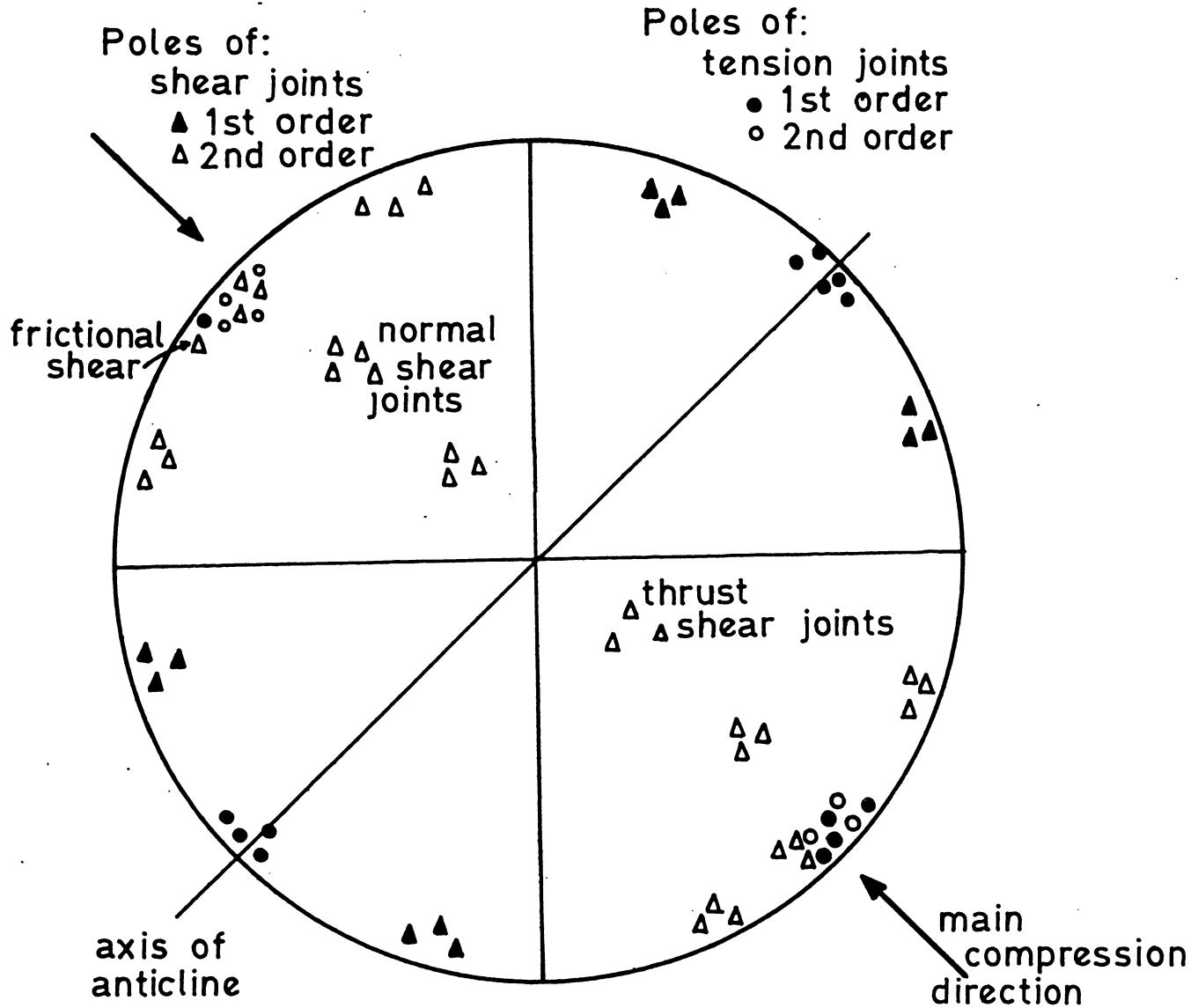
The joint poles were plotted on polar stereonets and studied in relation to shortening, assuming this to be the direction normal to the strike of the cleavage  $S_2$  or the bedding  $S_g$ , the latter in the cases where the  $S_2$  cleavage is poorly developed. The joint sets are described as dip, strike, or oblique if they trend parallel to the dip or the strike of the country rock, or are oblique to both.

Slip across joint fractures is rare. Dip as well as strike joints commonly have plumose structure on joint surfaces. Oblique joints are commonly clear cut.

### Theoretical Background

Circles imprinted on clay, when subjected to compression, elongate into ellipsoids by slipping on two shear surfaces oblique to the direction of shortening (Cloos 1930). According to Cloos the lesser the internal friction the greater the angle of shear, bisected by the axis of shortening.

Parker (1942) investigated the formation of high angled joints near the surface. He argued that vertical relief is easier than horizontal relief, and shear joints corresponding to reverse faults are more likely than vertical tension joints. He maintained that strike joints are high angled reverse faults inclined at 70-80° and trending normal to the compression direction. The attitudes depend on the following variables: 1. Elastic strain before rupturing.



**FIGURE 22. STEREOGRAPHIC REPRESENTATION OF POSSIBLE JOINTS IN AN ANTICLINAL STRUCTURE (after de Sitter 1964)**

2. Binding force. 3. Friction within rocks. Parker also noted the presence of subvertical oblique sets at approximately  $20^{\circ}$  to the dip direction of the beds.

De Sitter (1964, p. 99-111) recognized major first order sets of joints in folded rocks (i) the joints that trend parallel to the dip (tension joints), (ii) those trending parallel to the strike (shear and tension joints), (iii) a pair of shear joints oblique to the strike of the country rock. According to his stereographic representation the three major sets are vertical or subvertical (fig. 22). The oblique joints are analogous to the shear fractures obtained in the experiments by Cloos (1930). In addition, de Sitter visualised the possibility of normal shear and thrust shear joints.

In the fold structures, de Sitter (1964, p. 104) advocated an interplay of regional lateral compression and secondary effects within the fold structures. In an anticline there is tension across the fold axis in a zone above the neutral zone (assuming kinematically active bedding surfaces and concentric folding), and therefore the joint pattern is as follows in relation to the fold geometry: (i) dip joints in ac plane (normal to b-axis, formed by extension normal to the dip direction. (ii) Oblique joints with their obtuse angles of intersection bisected by the fold axis. (iii) Strike joints in the ab plane (parallel to axial plane), that trend parallel to the strike of the country rock, are formed due to elastic release normal to the fold axis.

A similar orientation of joint planes occurs in synclines, but the oblique joints form with the acute angle of intersection bisected by the compression axis. This difference arises as a compression occurs within the syncline adding to the effect of the regional compression, but below the neutral surface in a syncline would be the same as above the neutral surface in an anticline.

Price (1959 and 1966) considered that neither tension nor shear joints can develop during a compressive phase and that joints form because of lateral expansion and extension due to uplift. Rocks at depth are subjected to an overlying load, and being restricted, do not expand laterally. On uplift (i) a loss of vertical load by removal of rock by erosion, results in a change in magnitude of stresses in

a tri-axial residual stress system, (ii) uplift also allows the rock to occupy a larger area in the same angle of arc of the earth's sphere.

Sikander (1963) noted steep dip and oblique joints in Silurian slates in North Wales. The acute angle of intersection of oblique joints is bisected by the dip direction of the axial plane cleavage. The joint pattern in the overlying Carboniferous limestone and sandstone differs from that of the Silurian rocks in that it contains dip, strike, and oblique joints normal to bedding surfaces. The oblique sets intersect each other at an angle close to  $90^\circ$ .

### Joint Patterns in Lower Paleozoic Rocks

Cambrian-Ordovician Rocks. In the undivided Cambrian-Ordovician rocks, the rock exposures are poor and outcrops consist of roadside and river exposures. The joint data from each traverse were plotted in a central locality. The information was carefully collected at random on each exposure. Care was taken that a certain set was not repeatedly observed disproportionately to its actual occurrence. The joint pattern in the rocks is described for each major lithological group or a distinct rock unit.

Shickshock Group. The joint orientation varies with lithology (plate VIIa).

The metavolcanic rocks foliated by  $S_1$  are crossed by a strong dip set, with few joints in other sets. Locality SK 3 (plate VIIa) is a good example. In the diagram a maximum of a dip joint set poles coincides with the linear plot of fold plunges, that is the joints are at right angles to the fold axes. The relationship appears more consistent in the diagram than in individual outcrops, where slightly divergent axes are apparent.

The metasedimentary rocks foliated by  $S_1$  contain three major sets: a strong dip set (demonstrated by patterns of SK-6, SK-7, and SK-8), two oblique sets of unequal polar frequency maxima (SK-6, SK-7, and SK-8), and a steep strike joint set (SK-6). Some joints form normal to the  $S_8$  layering and  $S_1$  schistosity as in locality SK-7, but generally the joints tend to be steep, and lack the normal relationship to  $S_8$  and  $S_1$ . The acute angle of intersection of the

steep oblique joints is approximately bisected by the dip direction of  $S_g$  and  $S_1$  (plate VIIa, SK-6, SK-7, and SK-8).

The lack of normal relationship of the joints to the layering or schistosity indicates that the bedding and foliation anisotropy does not appear to have influenced the direction of local stresses responsible for the jointing.

Cambrian-Ordovician Slate. A joint pattern in the slate with  $S_2$  fractures transecting an earlier schistosity ( $S_1$ ) is demonstrated by a polar plot of joints at locality RM-1 (plate VIIa).

A majority of joint sets are vertical or subvertical and approximately in the dip direction. The maximum is elongated and poles to joints tend to lie in a great circle that contains both the average bedding  $S_g$  and cleavage  $S_2$ . That is, the joints tend to be normal to  $S_g$ . There is a secondary maximum of poles inclined at  $45^\circ$ , lying on the  $S_g$  great circle.

Ordovician Rocks. Most of data on joints in Ordovician rocks was collected from the well exposed coastal areas of the St. Lawrence River. The joint observations were taken at random, and care was taken to avoid being selective in observations. The patterns are described for each major rock type (plate VIIa).

Impure Grey Sandstone. The locality which is taken as typical for joints in these rocks is ME-7 (fig. 21). The pattern shows a simple relationship to  $S_g$  and  $S_2$  as shown in photograph 86 and plate VIIa. The pattern consists of a joint set trending slightly oblique to the dip direction of the layering  $S_g$ . Oblique joint sets intersect each other at an angle of nearly  $90^\circ$ . The acute angle of intersection is bisected by the dip direction. These joints are often accompanied by en echelon feather joints at about  $30^\circ$  to the oblique joint planes and normal to bedding (photo 82).

A faint joint set trends approximately parallel to the strike of  $S_g$ .

Most of the joints dip steeply, but some are nearly normal to bedding.

Calcareous Slates. Joint patterns in slates with minor calcareous siltstone and limestone are included. The slates contain a well defined dip joint set in relation to bedding (plate VIIa, ME 1, MW 5/6 (a) and MW 1).

Pairs of steep oblique joints are observed in localities MW 5/6/a and MW 4, and MW 1. In localities MW 5/6 (a) and probably MW 1, the acute angle of intersection between the oblique joints is bisected by the dip direction of the bedding.

In addition to the above described sets there are minor joint sets. In locality ME 1, the minor joints are normal to both bedding and cleavage surfaces, and dip at approximately  $45^{\circ}$ . The joint set in this locality is similar in orientation to that in locality RM 1 in Cambrian-Ordovician rocks.

In locality MW 9, a set of strike joint occurs normal to cleavage. In locality MW 5/6a, a minor strike joint set normal to  $S_2$  cleavage occurs in slate. In the minor limestone layers, however, there are oblique and dips joints normal to the bedding.

It is therefore observed that in the calcareous slates prominent sets of joints are usually steep in inclination and consist of dip joints and oblique joints, the acute angle of intersection of which is bisected by the dip direction of the mean bedding plane. The strike joints (as in locality MW 9) are rare but normal to the  $S_2$  cleavage surface.

Limestone and Quartzite. A limestone bed approximately 30 feet thick, enclosed in slate exposed in locality MW 5/6 (b) has a pattern different to that in slate at the same locality (MW 5/6a). As in the slate, a dip set is accompanied by oblique sets, and a strike set. The difference is that the joint planes are on average normal to the bedding.

In massive quartzite at locality ME 2 (plate VIIa) there are strong oblique sets and faint strike and dip sets. Joints are normal to bedding.

In addition to the patterns described above in the Cambrian and Ordovician rocks, there are irregular subhorizontal joints developed in slates at locality ME 1, and in the impure grey sandstone

(photo 84) at locality ME 9. The joint set is inclined at approximately 25°NE and it divides the cliff walls along the St. Lawrence River into irregular sheets two to three feet thick. The frequency increases upwards on the cliff walls.

Siluro-Devonian Rocks. Due to poor exposure of the Siluro-Devonian rocks, the limited joint data on each rock formation obtained from outcrops throughout the area is plotted in one diagram (plate VIIb). Figure 21 shows the distribution of the localities in each rock type investigated. The strike of bedding is reasonably constant throughout the units, allowing a broad division of the joints in the diagrams.

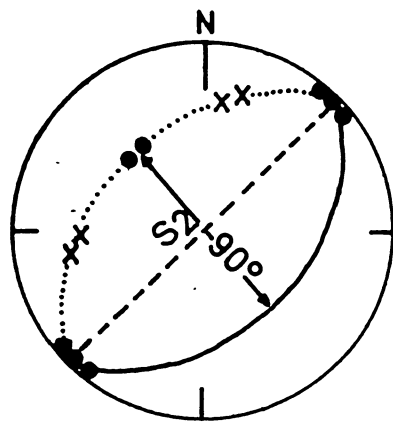
St. Leon Formation. The joint pattern in the Silurian rocks is best studied in the St. Leon Formation, which consists of interbedded shale, sandstone and siltstone. All lithologies contain dip joint sets. Strike joint sets and a few oblique joint sets occur only in the competent (sandstone and siltstone) units and are absent from the shale.

In the field it is apparent that the joints in the sandstone are normal to bedding and variable in angular relationship to the local fold plunges.

Cape Bon Ami Formation. The calcareous slates contain a dip set and a pair of oblique joint sets. The acute angle of intersection of the oblique sets is bisected by the dip direction of the  $S_2$  cleavage. The dip joints are vertical and normal to the mean fold plunge.

Grande Greve Formation. The pattern consists of closely spaced polar maxima formed in the northeast quadrant of the net. Both are considered as dip joint maxima in view of the large area represented in the diagram in which the strike of the rocks may vary.

York River-Heppel Formations. Dip joint sets predominate accompanied by one oblique set.



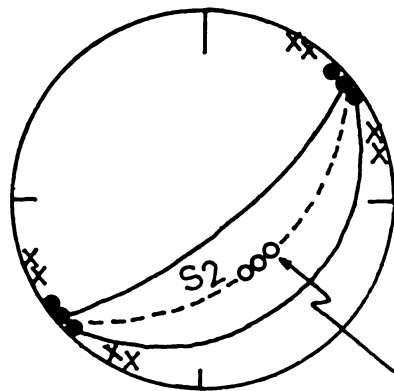
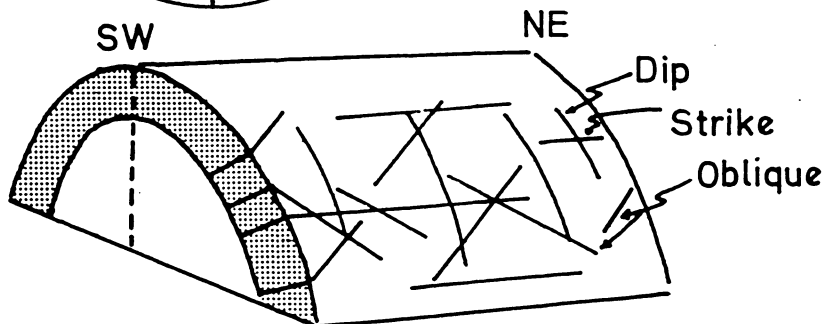
Tension Joint Poles (Dip and Strike)

- Frequent
- Relatively rare

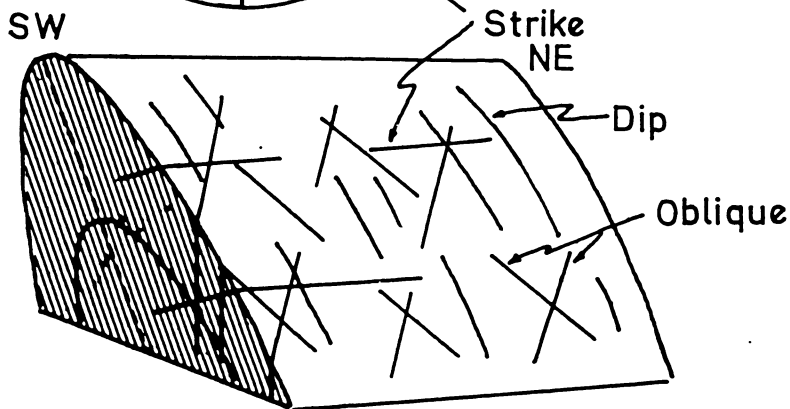
Shear Joint Poles (Oblique)

- x Frequent

a.



b.



**FIGURE 23. DIAGRAMMATIC JOINT PATTERN IN (a) COMPETENT WELL BEDDED LIMESTONE, SANDSTONE AND SILTSTONE, (b) WELL CLEAVED SLATE, SHOWN IN RELATION TO REGIONAL FOLDS WITH SUBHORIZONTAL AXES.**

Fortin Slate Group. The joint pattern consists of a single dip joint set.

Ste. Marguerite Volcanics. The joint pattern in the Ste. Marguerite volcanics consists of steep joint sets of variable strike. A major dip joint set maximum is formed with the poles at N 45E. The dip joints have the tendency to be northeasterly inclined. Two prominent oblique joint maxima develop in the eastern and southern parts of the net. Steep strike set maxima form in the SE and NW quadrants of the net.

### Joint Pattern and $F_2$ Fold Plunge.

Although in the patterns from calcareous slates at localities ME-1, MW-5/6 (plate VIIa), dip joints are common, these are variable in angular relationship to the local plunge of the folds. In the Fortin Slate Group (plate VIIb) the  $F_2$  folds plunge at all angles on the  $S_2$  surface, but the joints are consistently of the dip variety.

In sandstones and limestones similar relationships were apparent, but in massive beds the fold plunge is less variable, and hence the angular variation is not clear from a consideration of average attitudes as on nets.

### Conclusions

The joint patterns can be summarised as follows and deductions can be made about possible mechanisms:

(1) There is a definite relationship between joint orientation and lithology (fig. 23).

(i) Dip joints are common in all lithologies. They probably are tension joints in the area, parallel to a general direction of compression. Oblique joint sets are common within the slate and shale. They usually occur in pairs. The oblique joints are similar to the shear fractures obtained in the experiment of Cloos (1930). A consistent relationship between the fold structure and the

and the angle between oblique joints bisected by the bedding dip direction could not be established, but some joint patterns in Shickshock metasediments and Ordovician slates show an acute angle of intersection bisected by the compression direction (SK-6, SK-8, and MW 5/6).

(ii) In competent rocks, limestone and sandstone, dip and oblique joint sets are accompanied by strike joint sets. The strike joints probably formed as a set of tension joints perpendicular to the compression direction as a result of elastic release (de Sitter 1964). These joints are virtually absent in the incompetent rocks, but some slates are crossed by strike joints perpendicular to the cleavage.

(iii) In the competent and well bedded rocks most (dip, strike and oblique) joint sets are normal to bedding, whereas in the slates and the metavolcanics, although some sets are normal to layering, they are usually steep in inclination. They are also rarely normal to the cleavage. This is presumably due to the fact that in rocks containing sedimentary or tectonic anisotropy, compression is transmitted parallel to the anisotropy.

This relationship of the joint pattern to the lithology is similar to that observed by the worker in North Wales (Sikander 1963, also see p. 65).

(2) The dip joints are at variable angles to the plunges of local  $F_2$  folds, hence they cannot reliably be used as indicators of fold plunges, nor are they the result of extension parallel to fold plunges.

(3) The joints were probably formed as a result of residual compression after the main  $F_2$  folding phase, and they are not directly related to the kinematics of the  $F_2$  folds.

## 7. Faults

There are two major faults in the area, the Shickshock Fault and the Ste. Florence Fault. Both trend northeast-eastnorth-east and displace rocks subvertically by an unknown amount.

### The Shickshock Fault

The fault is marked by a wide zone of breccia and a valley along the southern margin of the Shickshock Mountains (fig. 6). Along Wilson Brook a tributary to the Cap Chat River, extensive breccia and dolomitic limestones that are not apparent elsewhere mark the Shickshock Fault zone. At this locality several fault slices bring Middle Ordovician and Silurian rocks against the older meta-volcanics to the north. The north side is the upthrown side.

Along the Matane River extensive outcrops of contorted and brecciated Upper Silurian St. Leon Formation are exposed on the south side of the valley. Often the talus slopes of breccia are so thick that they cover the true nature of the rock underneath. South of Matane River near Depot Jean the St. Leon Formation is brecciated to an almost unrecognisable rock.

Southwest of the Shickshock Mountain rocks along the fault zone are poorly exposed. Along the Tamagodi River the zone is not topographically conspicuous. Contorted and sheared rocks occur in the vicinity of the fault zone. Farther southwest, the fault follows lowlands into Matapedia Lake. Along the lake there is no visible evidence for the fault except for abrupt changes in stratigraphic units from the southern tip northward and from one side to the other. The branch fault marked as extending southwestward from the southeastern tip of Matapedia Lake is accompanied by no appreciable brecciation within the Siluro-Devonian rocks.

Truncation of the Lower and Middle Devonian rocks by the fault east of Matapedia Lake indicates an upper or post-Devonian age for some of the movements. The fault however, is considered to have been active more than once during the Lower Paleozoic.

Ollerenshaw (1963) regarded it as a conduit for the Cambrian

Shickshock volcanic rocks in the Shickshock area; and, because several ultrabasic bodies of Lower Ordovician age occur near or are cut off by the fault, he believed that the ultrabasics were also emplaced along the fault.

The Incognito Conglomerate-breccia at the base of the Middle Silurian succession contains fragments of the Shickshock volcanic and various other rocks (see Appendix I). The blocks of up to 2 feet diameter were probably contributed by erosion of the volcanics and meta-arkosic rocks from a fault escarpment during the Middle Silurian.

The presence of outcrops of Middle Silurian Val Brilliant quartzites overlying the Cambrian-Ordovician rocks north of the Shickshock Fault (Ollerenshaw 1961, fig. 6) brings to light several other considerations.

(i) The Val Brilliant quartzites also occur south of the Shickshock Fault south of Mount Albert, east of the area. The quartzites are mature and occur with remarkable lithological uniformity on both sides of the fault. This indicates that they were deposited on a surface of low relief.

(ii) The Siluro-Devonian rocks alone, between the Shickshock and Ste. Florence faults, attain a total average thickness of 25,000 feet. The youngest rock type of the York River-Heppel Formations is brought up in contact with the Cambrian-Ordovician and probably Cambrian slates along the south tip of Matapedia Lake. Therefore to account for the pattern the Shickshock Fault presumably has a throw of more than 25,000 feet.

During the deposition of the lowest Middle Silurian beds escarp forming displacements gave rise to conglomerate-breccia. Then there was pene-planation followed by deposition of the quartzite. This was followed again by movement in Upper or post-Devonian times which brought about the present juxtaposition of the rocks of various ages and brecciation of Upper Silurian and Devonian rocks.

The displacement on the Shickshock Fault resulted in a northward tilt in the Shickshock block analogous to a tilt on a trap door.

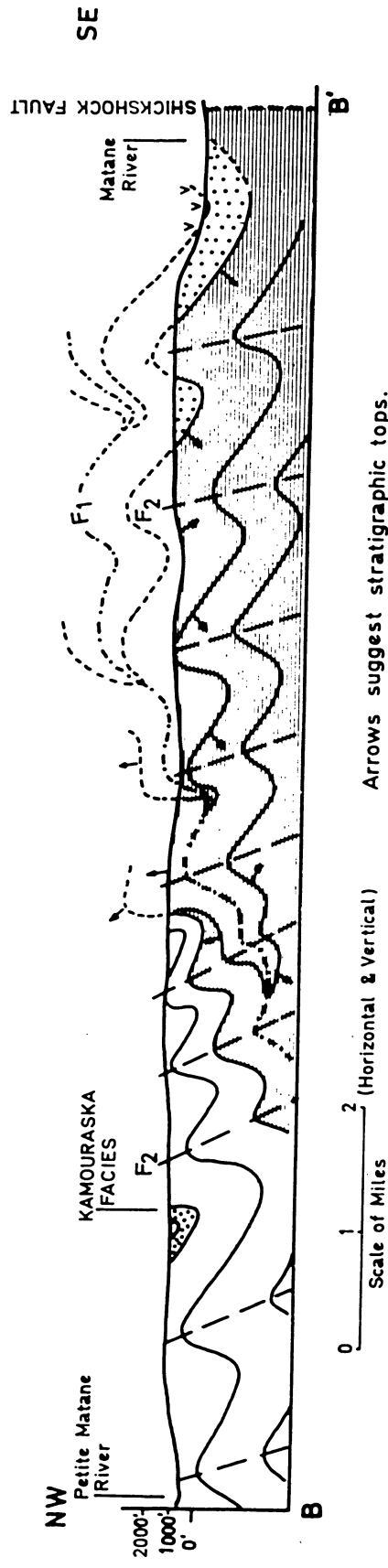
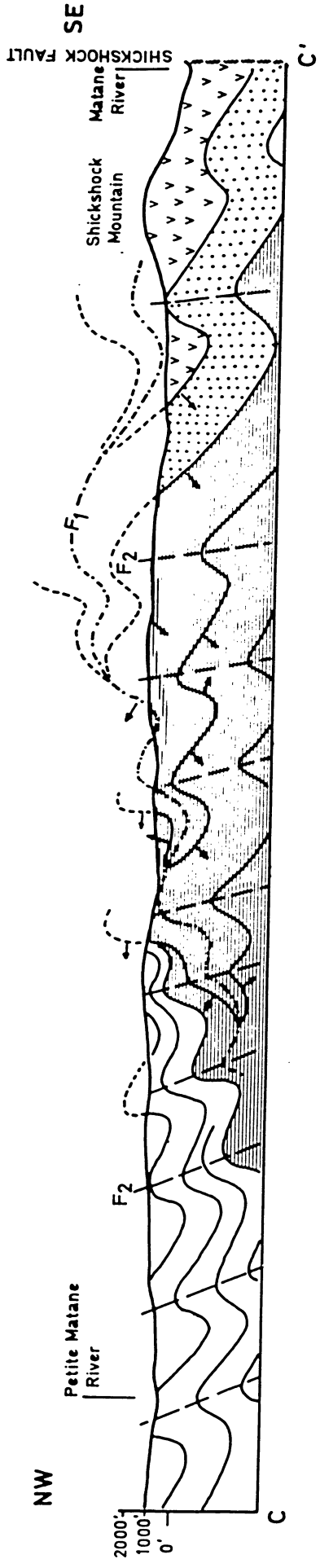
Other examples of this asymmetric 'trap door' uplift are the near vertical faults in the Wyoming Rocky Mountain (Eardley 1962, p. 361-374), the faulting between the Franciscan and Granite basements in California (Reed, 1951, p. 284), and asymmetric wedge uplift of the basement in the Saudi Arabian oil fields (Powers et al. 1963, verbal communication F.K. North 1966).

### "Ste. Florence" Fault

This fault trends N70E and traverses the area just north of Ste. Florence (fig. 6). It separates the Middle and/or Lower Devonian Fortin Group on the south from calcareous sandstone of the Middle Devonian York River-Heppel Formations on the north. The fault is marked in the Fortin Group by an increase in micaceous minerals and by contortions of the Fortin slates (Stearn 1959), and in the Heppel Formation by folds. The rocks on the two sides are of different facies but of little difference in age. Thus the displacement on the fault surface is difficult to estimate.

Smaller shear zones that trend parallel to the strike of the layering are marked by sheared rocks, especially within the calcareous shales of the Ordovician Quebec Group, but these are not regarded as having large-scale displacements. Usually these shear zones mark the contacts between rocks of differing competency. They are common along the contact of shale with limestone, sandstone or quartzite.

- ORDOVICIAN SLATE
- CAMBRIAN-ORDOVICIAN SLATE
- F<sub>2</sub> AXIAL PLANE
- ▨ SHICKSHOCK METASEDIMENTARY ROCKS
- ▨ SHICKSHOCK METAVOLCANIC ROCKS
- F<sub>1</sub> AXIAL PLANE



**FIGURE 24.** STRUCTURAL INTERPRETATION OF THE SHICKSHOCK MOUNTAIN AREA. SECTIONS DRAWN ALONG C-C' AND B-B' AS SHOWN IN THE GEOLOGICAL MAP, FIGURE 6.

## 8. Structure of the Shickshock Mountain Area

In an interpretation of the structure of the area the following observations are relevant:

(i) The  $S_1$  schistosity in the metavolcanic rocks and in the matrix of the interbedded metasediments, and also in the structurally underlying sediments is tectonic and closely parallel to  $S_s$  indicating isoclinal folding.

(ii) The  $S_1$  schistosity is not restricted to the Shickshock Group, it extends into the belt of Cambrian slates lying to the north of the Shickshock Mountains.

(iii) A Middle-Upper Cambrian K-Ar age, of  $530 \pm$  m.y. (Ollerenshaw 1963), dates the development of schistosity in the Shickshock Group, and implies that the rocks are older.

(iv) The Mount Albert igneous body is considered as intruded into schistose metavolcanic rocks. The K-Ar age of the contact aureole is dated at  $495 \pm 35$  m.y. This again indicates a pre-Lower Ordovician age for the Shickshock Group (MacGregor 1962).

(v) The schistosity is folded into broad upright and steeply inclined  $F_2$  synclines and anticlines.

(vi) Due to longer southeasterly limbs of the  $F_2$  anticlines the general dip of the layering  $S_s$  and schistosity  $S_1$  is southeasterly in the Shickshock Mountains.

(vii) The metavolcanics, metasediments and the grey slate to the north intergrade. Therefore, no thrust contacts are invoked within this conformable sequence. The rocks north of the sequence are paleontologically dated as Lower Ordovician (in the Tourelle and Courcellette areas, 55 miles northeast of Matane, McGerrigle 1954). Therefore the southeasterly dipping rocks are probably inverted and young northwestward.

An interpretation of the structure of the Shickshock Mountains is given in fig. 24. The metavolcanics form the core of an  $F_1$  isoclinal recumbent anticline that has been refolded by  $F_2$  folds. The meta-arkose and other metasediments that structurally underlie the metavolcanics are younger in age. These are structurally underlain to the north by the younger Cambrian-Ordovician grey slates. In this

belt on the inverted limb of the  $F_1$  fold the rocks dominantly dip southward and young northward, contrary to the area lying further north where the rocks are Ordovician in age and dominantly dip and young southward.

The isoclinally folded Cambrian-Ordovician rocks make up the basement for the Ordovician rocks which contain structures assignable to only one phase of deformation.

The isoclinal and recumbent nature of the  $F_1$  folds render the rocks deformed by  $F_1$  and  $F_2$  difficult to distinguish from the rocks deformed solely by  $F_2$ . Before  $F_2$  deformation, the horizontal layering,  $S_s$ , and the  $S_1$  schistosity were virtually parallel to the undisturbed bedding of the overlying rocks, and hence no discordance is apparent between  $S_s$  layering affected by  $F_1$ ,  $S_1$ , and beds not affected by  $F_1$  folding.

## IV CONCLUSIONS

1. Summary and Interpretation

Table 3 (p. 26) summarises the essential characters of the various generations of folds in the area.

The  $F_1$  folds are isoclinal and recumbent. The axial plane schistosity  $S_1$  is characterised by a parallel alignment of micaceous minerals. The schistosity  $S_1$ , restricted to the Shickshock Group and grey slates of Cambrian-Ordovician age, is considered as tectonic mainly because: (i) It develops in all lithologies and contains rotated minerals and kinematic plagioclase. (ii) In most localities an axial lineation due to mineral elongation or  $S_s/S_1$  intersection is present. (iii) In a few localities schistosity  $S_1$  and layering  $S_s$  can be seen to diverge, indicating proximity to  $F_1$  fold hinges. The axial lineations are restricted to a belt of rocks containing the  $F_1$  structures that extends southwest of Matapedia Lake to an inlier of grey slate intruded by an ultrabasic igneous body.

The oldest rocks that do not contain the  $F_1$  structures are Lower Ordovician in age. These are impure grey sandstones exposed near Ste. Felicite and the Tamagodi Sandstone near the southwestern edge of the Shickshock Group. This is consistent with radiogenic dates that time the  $F_1$  deformation as Middle-Upper Cambrian. Within the Shickshock fault zone a wedge of Middle Ordovician slates dated by graptolites is not deformed by  $F_1$ .

A line of demarcation, separating the Cambrian-Ordovician rocks containing the  $F_1$  and  $F_2$  structures from rocks containing only the  $F_2$  structures also separates the undivided Cambrian-Ordovician from the Ordovician rocks (fig. 6). The  $S_1$  schistosity was probably subhorizontal before deposition of Ordovician and later rocks and  $F_2$  folding along the same axes of folding as  $F_1$ . This parallelism between the  $S_1$  schistosity in the Cambrian-Ordovician and bedding in the Ordovician makes it extremely difficult to distinguish between  $F_2$  folds in the Ordovician and later rocks and  $F_2$  folds that re-fold  $F_1$  in the older rocks.

The presence of the pre-Middle Ordovician deformation is

better demonstrated east of the area where the  $S_1$  schistosity and  $L_1$  lineation are oblique to later  $S_2$  and  $L_2$ .

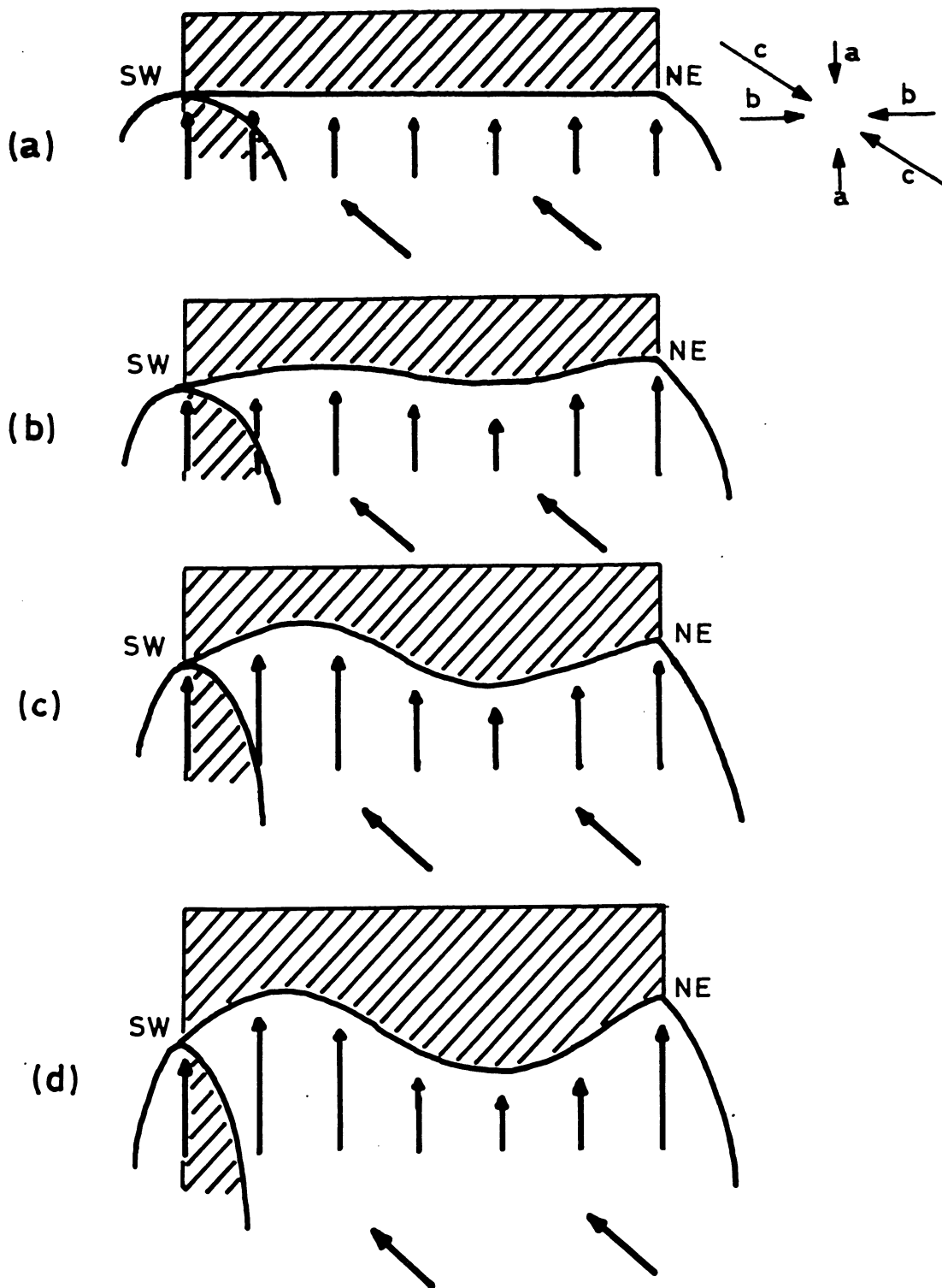
Regional metamorphism within the Cambrian-Ordovician rocks is attributed to the  $F_1$  phase of folding. The isograds drawn by Mattinson (1964) are oblique to the  $F_2$  fold trends. They also cut across the contact between the Shickshock Group and grey slates. The regional metamorphism is not considered to be related to granitic activity (as invoked by Mattinson) but to metamorphism accompanying the  $F_1$  folding.

The youngest rocks involved in the  $F_2$  folding are Middle Devonian. The  $F_2$  folds deform all Lower Paleozoic rocks to form the Gaspé Synclinorium, the axial trace of which varies between northeast and eastnortheast.

The  $F_2$  folds are parallel and transitional to each other in their trends and inclinations of axial planes and no evidence has been found for a phase of intense folding within the Middle Ordovician-Silurian-Devonian succession prior to the main phase of  $F_2$  deformation. In the southern part of the area near Chaleur Bay, a complete transition occurs among Ordovician-Silurian and Devonian sediments. However, a phase of non-deposition may have occurred between the Upper Ordovician and Lower Silurian in the northern part of the area. For example, in the vicinity of Matapedia Lake and to the southwest, the oldest Silurian beds are Middle Silurian in age, and Upper Ordovician beds are not found. The Upper Ordovician-Lower Silurian unconformity is probably a manifestation of an uplift and gentle warping of the older sediments.

In view of the variable tightness of the folds, it is concluded that this is no criterion by itself to attribute the folds to different ages. The varying acuteness of  $F_2$  folds in the rocks as demonstrated in the Causapsal area (p. 15) is partly a function of lithology (also shown experimentally by Ramberg, 1961).

Thus, parallel structures of different ages can not be differentiated on the basis of fold style alone. The Radiogenic dating of recrystallised micas in crenulation cleavage should date this structure (Rickard 1964). The worker, following Rickard, also considers the possibility of 'tectonic peninsulas' or 'tectonic islands' of older tectonic age existing within rocks with a younger



**FIGURE 25.** DIAGRAMMATIC GROWTH OF F<sub>2</sub> FOLDS WITH VARIABLE AXIAL PLUNGE AS VISUALISED IN SUCCESSIVE STAGES OF DEVELOPMENT. THE COMPRESSION IS CONTINUOUS NORMAL TO THE AXIAL PLANES.

tectonic history that have folds which are indistinguishable in style and orientation.

$F_2$  folds vary in style from open and conjugate to isoclinal. Competent layers, limestone, siltstone, and sandstone tend to be of uniform thickness as with concentric folding, whereas the incompetent slates are largely similarly deformed. The ideal bedding thickness relationships relating the folding to either of the two mechanisms are not strictly adhered to in the area. The folds often show a thinning of the competent layers on the limbs. Rarely, also, the hinge area of the competent layer may fracture and blocks of it be punched outwards.

The ratio of wave lengths of the folds to the thickness of competent layers is approximately constant (between 1:4 - 4.4), and indicates a maximum shortening of the competent layers by concentric folding. The folds were, thus, initiated by a concentric mechanism, whereby the competent layers transmitted compression and bedding surfaces were kinematically active. Similar folding and flattening became dominant in the later part of folding when further concentric folding was no longer possible. During this phase the competent layers behaved passively and yielded to the flow within incompetent beds, giving rise to similar thickness relationships.

The  $F_2$  fold axes lie on the axial planes at all angles while the slickensides on the calcite veins maintain a predominant direction parallel to the dip. It is considered that the slickensides on the calcite vein parallel to the axial planes are due to a movement of material parallel to the axial planes during the later modification of the  $F_2$  folds by the similar mechanism.

The accentuation of the concentric folds by similar folding and flattening consisted in differential advance of portions of the folds. These accentuated folds maintained the shortened but concentrically imposed fold wave lengths. The cleavage planes provided the flow planes, and separated the different domains in folds that are characterised by differing amounts of advance (Carey 1962, p. 98-99). A variation in the advance of the different parts of the folds along the trace of the axis, gave rise to the variable plunge of the folds (fig. 25).

The  $F_2$  folds are thus doubly plunging. The plunges are from  $0^\circ$  to  $90^\circ$  in either southwest or northeast directions. In three dimensions a competent layer conforms to an advancing incompetent 'front' which resembles an elongated dome, somewhat like an upturned canoe. Such canoe shaped outcrop patterns are common in the calcareous slates of the Quebec Group along the coast of St. Lawrence. The S and Z-shaped outcrops depend on the planar sections of the plunging ends of the folds; the southwesterly plunging end of the folds gives rise to a Z-shaped outcrop pattern, and northeast an S-shaped plan in the northern part of the area where folds are overturned northwestward.

The development of folds first by concentric, and then by similar folding and flattening is considered a continuous process of deformation which involved a shortening in a northwest-southeast direction. A triaxial compression system would have the greatest compression axis in a NW-SE direction; intermediate compression axis in NE-SW direction, and the smallest compression axis in a plane normal to the plane containing the other two axes, and parallel to the mean of the axial plane of the  $F_2$  fold.

The axial plane fracture cleavage,  $S_2$ , consists of a fracture set, which when extremely well developed resembles slaty cleavage with a weak mineral alignment parallel to the cleavage. The  $S_2$  cleavage is classified as shown in table 4. It is suggested that when the fracture cleavage displaces markers less than  $1/4$  that of the fracture spacing, ( $d/D < .25$ ), it should be classed as fracture cleavage, and when the displacement across the fracture is  $1/4$  or more than that of the average spacing of the cleavage ( $d/D > .25$ ), it should be termed as fracture-slip cleavage. The axial-plane fracture cleavage is called crenulation-slip cleavage when there is an early schistose fabric or a pronounced sedimentary anisotropy crenulated with  $d/D > .25$  across the cleavage surfaces. The crenulation-slip is most obvious in the rocks with pronounced marker surfaces. In homogenous rocks the folding deformation between cleavage planes is less obvious. It is, however, indicated as a bulge between the two cleavage traces on bedding surfaces.

The spacing of cleavage depends on the rock type and the angle of its intersection with bedding. In a given lithology the

cleavage is closely spaced when the angle  $S_1/S_2$  is small, and wider in spacing when the angle is large. The cleavage therefore is not uniformly developed. It forms as bands of widely and closely spaced cleavage, along the hinge and limb areas respectively.

The lack of the intense cleavage growth in the hinge areas of the folds is regarded as due to the following: (i) There is less amount of slip across cleavage involved in the hinge areas of the folds, and therefore it is accomplished by movement along fewer cleavage planes. (ii) The presence of thickened "caps" of competent layers in hinge areas reduces the frequency of the cleavage. The thickened cap is more competent than the surrounding rock, therefore less susceptible to flattening, and hence the cleavage in the hinge area is widely spaced as compared to that on the limbs where rock has been flattened and the distance between the cleavage planes reduced.

The  $F_2$  folds are overturned to the northwest in the northern area, to the southeast in the southern area, and upright in the central part of the area. The area is divided into two major domains: I, contains shorter limbs to the north in northwesterly overturned anticlines; II, contains shorter limbs to the south in southeasterly overturned anticlines. Domain I can be subdivided into: (i) a northern area with inclined  $F_2$  axial planes; (ii) the southern part in which the axial planes of the  $F_2$  folds are upright or subvertical. The Shickshock Fault separates the two subdomains.

The inclination of the  $F_2$  axial planes is closely related to the symmetry of profile and limb ratio of the  $F_2$  folds. The regional orientation of the axial plane cleavage, therefore, is associated with the development of the  $F_2$  folds in the area. Thus, the Gaspé cleavage fan is a result of development of regional folds with upward diverging axial planes and not due to formation of upright folds and a later fanning. However, an increased overturning of the fold axial planes along the Shickshock Fault, and the inclinations of  $45-50^\circ$  in the axial planes in the northern Ordovician rock belt as opposed to steep axial planes in the southern part of the area, are tentatively considered as due to a later modification of the axial plane attitudes.

To the north of the trace of the Shickshock Fault there is a sharp decrease in the dip of the  $F_2$  axial planes. Originally nearly

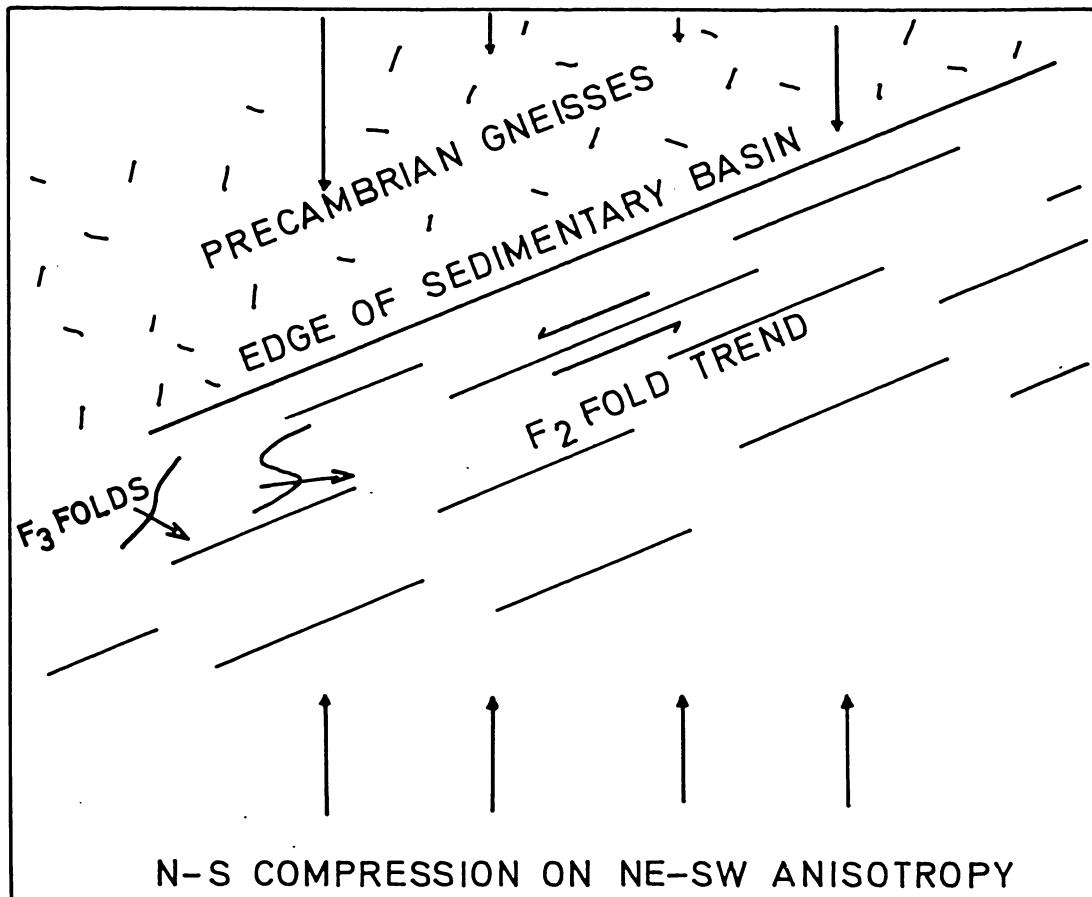


FIGURE 26. DIAGRAMMATIC EXPLANATION TO ACCOUNT FOR THE NE-SW ANTICLOCKWISE COUPLE AND THE F<sub>3</sub> FOLDS IN THE NORTHERN PART OF THE AREA.

upright  $F_2$  folds were rendered inclined in the schistose Cambrian-Ordovician rocks by northerly tilting of the Shickshock fault block. Farther north, the rate of change of dip is more gradual, and the folds are more overturned. It is possible that the slope of the northward tilted Shickshock block gave rise to a sliding of the folded Ordovician sediments under gravity. This may have resulted in a maximum compression axis inclined to the northwest, and consequently a greater northwesterly overturning of the folds.

$F_3$  folds mesoscopic and macroscopic in scale are restricted to the St. Lawrence coast, and are not found in the interior. They are commonly S-shaped and deform the  $F_2$  structures. The folds are considered as genetically related to subhorizontal slickensides on calcite veins parallel to the  $F_2$  axial cleavage planes and sinistral displacements along the cleavage planes. In view of the fact that the  $F_3$  folds deform the  $F_2$  folds into acute S forms (fig. 18) they are considered as formed in conjunction with a NE-SW anticlockwise couple acting along the NE-SW anisotropy formed by the  $F_2$  folds. This couple can be conceived as due to a N-S compression directed obliquely across the anisotropy. Since the  $F_3$  folds are restricted to the northern part of the area, which is closer to the edge of the sedimentary basin, it is possible that the compression could result from a northward push against the edge of the basin (fig. 26).

The  $F_4$  folds are restricted to the rocks containing strong  $S_1$  anisotropy, and are best developed in rocks of the Shickshock Group.  $F_4$  folds are microscopic and mesoscopic open angular symmetric or asymmetric kink folds trending variably north or northwest. They are S and Z-shaped and are regarded as pairs of conjugate folds.

The  $F_4$  folds deform the  $S_1$  schistosity. The axial surfaces are planar across  $F_2$  folds, therefore  $F_4$  are younger than  $F_2$ . The relative age between  $F_3$  and  $F_4$  is not definite but  $F_4$  is suggested as being younger. They are considered as formed by a cross folding of  $F_1$  and  $F_2$  structures by compression in a NE-SW direction parallel to the  $F_1$  and  $F_2$  structural trend. (fig. 27d).

Kink bands with low dipping axial planes in the northern limb displace the upper rocks away from the axial area of the Gaspé Synclinorium. The kink bands are associated with low northwesterly

dipping slip planes and fine low plunging microkinks with displacements similar to that of the larger kinks. The kink bands are not observed in the southern part of the area, presumably due to a lack of well cleaved rock type.

The kink bands, the associated weakly developed cleavage and micro-kinks are considered as due to post-tectonic residual stresses. It is considered that on removal of the northwest-southeast directed lateral stress that acted during the main phase of  $F_2$  compression, the indurated rocks expanded away from the axial area, the upper parts moving farther. Away from the axial area in the north, displacements took place in a uniform northerly direction, but along the axis and slightly to the south, displacements in both northerly and southerly directions occurred. Owing to the presence of well cleaved fine grained non-calcareous rocks these movements are only indicated along the axial area and in the northern limb of the Gaspé Synclinorium.

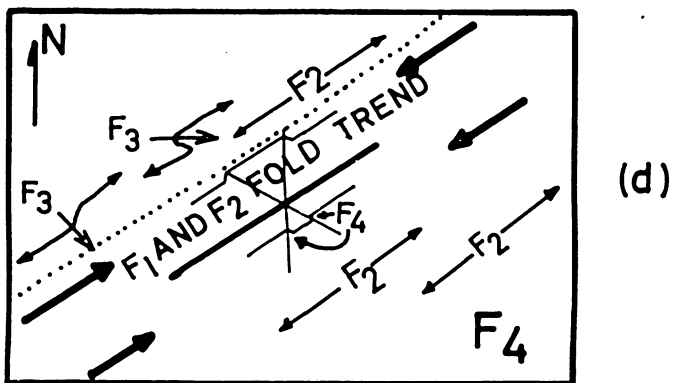
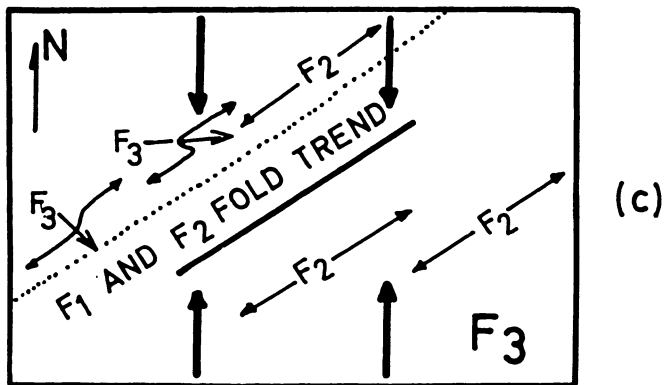
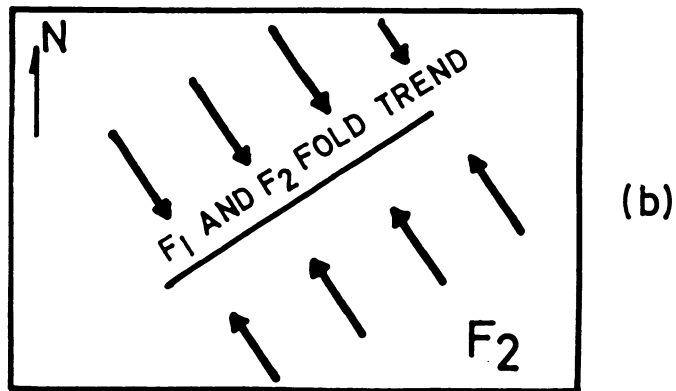
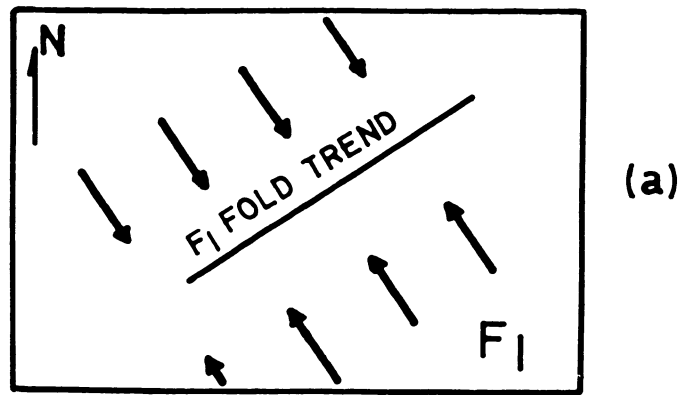
Farther south of the synclinorium axis the kinks were not observed in the Fortin Group slates. To the south of the Fortin slate belt, Upper Ordovician Matapédia group slates are calcareous, and Siluro-Devonian rocks are calcareous and volcanic. It is considered that largely due to an absence of a suitable lithology, the deformation which presumably consisted of upper rocks moving south, is not preserved in the rocks.

Joints in competent beds (limestone, siltstone, and sandstone) and in incompetent slates are in dip and oblique sets formed as extension and shear fractures. In the competent beds the joints are perpendicular to the layering, and in the incompetent beds they are usually steep with no consistent relationship to layering. In addition, in competent rocks there are strike joints apparently formed as elastic release fractures (de Sitter 1964), which are usually absent from incompetent slates. Some slaty rocks, however, contain strike joints perpendicular to the cleavage anisotropy. En echelon feather joints are formed along oblique joint sets and indicate shear directions across the oblique fractures.

The joints bear no direct relationship to the plunge of the  $F_2$  folds. This is true especially in the incompetent argillaceous rocks in which steep dip joints are consistent, and the fold axes plunge at

variable angles. Therefore, the joints are structures formed as due to regional stress readjustments and not as a result of extension in a direction parallel to fold axes.

The joints are considered as formed after the lateral compressive forces that gave rise to the  $F_2$  folds died out by a release of residual energy in the rocks and its interplay with gravitational loading during uplift (Price 1966, p. 135). The residual stresses were presumably transmitted along the anisotropy, particularly the bedding in the competent rocks and occasionally along cleavage, to give rise to the commonly observed perpendicular relationship of the joints to the layering or cleavage.



**FIGURE 27.** DIAGRAMMATIC PLAN OF THE DIRECTION OF MAXIMUM COMPRESSION IN EACH TECTONIC PHASE IN RELATION TO FOLD TRENDS. **F3** FOLDS OCCUR NORTHWEST OF THE DOTTED LINE.

## 2. Tectonic History of the Area

The following sequence of tectonic events is suggested for the western Gaspé Area:

1. A tectonically active northeast-southwest basin in which shales and coarse clastics were deposited with continued volcanic activity was severely folded by NW-SE compression in late Cambrian-early Ordovician times (fig. 27a). The deformation gave rise to NE trending recumbent isoclinal folds and parallelism between bedding  $S_s$  and schistosity, a  $S_1$  mineral elongation, and an increased metamorphic grade.
2. Ultrabasic rocks intruded the schistose rocks in Late Cambrian-Early Ordovician times.
3. Lower-Middle Ordovician rocks were deposited on sub-horizontal  $S_1$  schistosity in the northern part of the area. In the southern part continuous deposition took place from Ordovician to Devonian times.
4. A phase of uplift and erosion and presumably slight warping took place in Upper Ordovician-Lower Silurian times in the northern part of the area. Movement along the Shickshock Fault probably also occurred in this period, closely followed by a phase of planation of the fault escarpment.
5. A period of basic volcanic activity took place in Lower-Middle Devonian times in the southern part of the area.
6. Renewed NW-SE compression (fig. 27b) and post Middle Devonian (Acadian) deformation of the Ordovician-Silurian-Devonian rocks gave rise to northeast trending folds ( $F_2$ ) parallel to  $F_1$  and of variable tightness throughout the area. Renewed movement along the Shickshock Fault occurred towards the end of the tectonic phase, and gravity sliding of Ordovician sediments probably took place on a northwesterly inclined Cambrian-Ordovician basement.
7. A phase of deformation in the northern area resulted in acute S-shaped  $F_3$  folds (fig. 27c) in the St. Lawrence coastal area. This was probably accompanied by a sinistral couple acting in NE-SW direction along the Acadian structures. The

$F_3$  folding and sinistral couple could have resulted from a N-S compression.

8. NE-SW compression and conjugate folding  $F_4$  (fig. 27d) occurred in the Cambrian-Ordovician rocks with a closely spaced and well developed  $S_1$  anisotropy.

9. A general relaxation of the compression took place towards the end of the tectonic history giving rise to subhorizontal kink bands, crenulation lineation, micro-kinks and joints.

### 3. Comparison of Tectonic History with Other Areas in Northern Appalachians

Ayrton (1964) in a study of the Chandler-Port Daniel area in eastern Gaspé invoked a Gaspesian Orogeny (pre-Middle Ordovician) resulting in tight northwestward overturned folds and axial plane schistosity in greywacke, volcanics, phyllitic siltstone and quartzite of the Macquereau Group, which may be of Precambrian age. This was followed by a phase of ultrabasic intrusion, probably as a manifestation of a Taconic (Upper Ordovician-Lower Silurian) orogenic pulse, and then an extensive Acadian (Devonian) deformation. The last mentioned deformation had little effect on the Macquereau group of rocks. The Acadian structures in Ordovician and younger rocks consist of eastnortheast trending open upright folds and axial plane fracture cleavage.

Although Ayrton (p. 84), considers the possibility of folding of Middle Ordovician rocks by Taconic as well as Acadian orogenies, he favours the view that the Middle Ordovician rocks contain only the Acadian deformation structures (p. 85). He rejects the suggestion of Skidmore (1958, p. 6) that the cleavage in the Middle Ordovician rocks was formed as a Taconic structure and then folded by the Acadian deformation.

The two deformations invoked by Ayrton can be correlated in age and style of folds with the  $F_1$  and  $F_2$  phases of folding in western Gaspé.

In the Knowlton-Richmond area in southern Quebec, Osberg (1965) suggested at least two phases of deformation in Cambrian metasediments and schists. The early isoclinal folds have been preserved in the flanks of slightly oblique late folds. The linear elements of the late folds vary in pitch on a plane defined by the axial plane of the corresponding folds. The age of the early phase may be post-Ordovician and pre-Silurian, and that of the later phase may be Upper Devonian. The age relationships are ambiguous but derived from stratigraphic considerations of the adjacent areas. The early deformation in the Knowlton-Richmond area can be related to the  $F_1$  phase in western Gaspé/<sup>in</sup>fold style, but not in age. The late deformation is apparently related to the  $F_2$  folding in style of folding, and is slightly younger in age.

de Romer (1961), considering the St. Etienne-de-Bolton area in southeastern Quebec distinguishes two major styles of folds; east-west

trending tight to isoclinal folds and superimposed on them, north-south trending relatively open folds.

The early structures consist of small scale isoclinal folds in slaty, phyllitic and schistose rocks. A strong axial plane schistosity is characterised by re-orientation of micas; an early lineation consists of pods of quartz which are hinges of early folds. The late structures consist of open folds which deform the early schistosity, and contain axial plane fracture and slip cleavage. A late lineation is defined by the intersection of early schistosity or bedding and late cleavage.

There are two phases of metamorphism which are related to the two phases of deformation, and both are para- and post-kinematic.

The early phase of deformation affects the middle Ordovician Beauceville Group (p. 280) and older rocks, whereas the late phase of deformation includes Siluro-Devonian rocks. de Romer regards the two phases of folding as representing a long continued period of deformation.

In western Gaspé there is probably no deformation equivalent in age to the late Ordovician folding distinguished by de Romer in southeastern Quebec. The late deformation which folds Siluro-Devonian rocks is equivalent in fold style and age to the  $F_2$  folding in western Gaspé.

Fyson (1966) distinguishes three phases of folding in the early Paleozoic Meguma Group of Nova Scotia. Early middle Devonian main folds,  $F_1$ , are upright, low plunging, trend northeast to east, and contain upright axial plane cleavage. The  $F_2$  folds are angular and Z-shaped in profile, have steep axial planes striking north and to the northeast and are older than mid-Devonian granites. The  $F_3$  folds are open, S-shaped kinks with steep axial planes striking northwest, and are post the mid-Devonian granite intrusion. The main folds were formed by a northwest to north compression perpendicular to the  $F_1$  axes, which changed to an east-west direction, oblique to the  $F_1$  fold trend, and gave rise to the  $F_2$  and  $F_3$  folds.

The Middle Devonian  $F_1$  folding of the Meguma Group can be correlated in age and style of folds with the  $F_2$  deformation in the western Gaspé area.

Although a different mechanism has been proposed, the  $F_2$  folds in the Meguma group of rocks may be equivalent in age to the  $F_3$  folds in the western Gaspé area on the following basis: The  $F_2$  folding in the Meguma rocks as with the  $F_3$  folding in the western Gaspé area, closely followed the main middle Devonian deformation. The commonly occurring Z-shape of  $F_2$  folds of the Meguma rocks can be related to the rare Z-shape of the  $F_3$  folds in the western Gaspé area. In the Meguma rocks rare S-shaped  $F_2$  folds form as conjugate pairs with Z-shaped folds; similarly, the rare Z-shaped  $F_3$  folds in the western Gaspé presumably form as conjugate pairs with adjacent S-shaped  $F_3$  folds.

The  $F_3$  folding of the Meguma rocks is probably equivalent to the  $F_4$  phase of deformation in the western Gaspé area, the essential difference being, whereas the cross folding of the Meguma rocks was due to a compression oblique to the main fold trend, thus forming consistently S-shaped kinks; in the western Gaspé area the compression was parallel to the main fold trend forming conjugate S- and Z-shaped kinks.

## V SUGGESTION FOR FUTURE WORK

In the present study more detailed work on the  $F_1$  structures was limited in view of the large size of the area and the limited time. The  $F_1$  structures and their age and geological mapping require careful attention. In this respect a small area including the northern part of the Shickshock Mountains and the rocks to the north should be further examined. For such a study the Mount Albert Area, where  $F_1$  and  $F_2$  structures diverge on a regional scale, is recommended.

The overall structure of the Shickshock Mountains is far from being satisfactorily interpreted.

Conodonts may yet prove to be the most satisfactory means to correlate the Ordovician slates in the area, especially where the cleavage has destroyed other fossil records. For such a study to help solve the regional structure, the limestone bands in the slates should be systematically sampled for conodonts.

Radiogenic dating of recrystallised micas in the slates may provide clues to the age of deformation in the area.

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## VII APPENDICES

APPENDIX ISTRATIGRAPHY AND LITHOLOGICAL DESCRIPTION OF ROCK TYPES IN THE AREA:<sup>1.</sup>

## CAMBRIAN ORDOVICIAN ROCKS:

These rocks crop out exclusively in a belt lying between the St. Lawrence coast and the Shickshock Fault, which extends northeast along the southern margin of the Shickshock Mountains.

The rocks consist of dark grey slate, sericitic or phyllitic, compact when silty and thickly bedded. They contain interbeds of light coloured siltstones and limestone in dark grey calcareous slate. These rock-types occur in association with quartzite and limestone with local development of breccia and conglomerate.

The sequence of fine and coarse grained clastics of Cambrian-Ordovician age contain massive interbeds of green schistose metavolcanic and associated arkosic metasedimentary rocks (Ollerenshaw, 1963). These comprise the Shickshock Group. The slates exposed immediately to the north of the Shickshock Mountain and extending in a parallel trend east of the Matapedia Lake are silvery in lustre and are highly sericitised. These slates, complexly folded, grey phyllitic, stained reddish on the surface, and interbedded with thin grey silty layers, are also exposed southwest of the Shickshock Mountains. The schistosity is cut across by a younger strain slip cleavage. These rocks are Cambrian- and probably Lower Ordovician in age.

Limestone is poorly developed as thick massive beds. These are foliated but do not exhibit the schistosity demonstrated by fine lithologies.

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1. Dimensions are given in metric scale in microscopic study.

SHICKSHOCK GROUP: It consists of the following rock units:

Sandstone: The sandstone is regarded by Mattinson (1958) as a part of the Quebec Group rather than the Shickshock Group in which it is included by Ollerenshaw (1963). The rock is usually greenish grey, weathering light grey in colour. It occurs along the southwestern edge and the northern margin of the Shickshock Mountains with general southeastward dip of the Shickshock Group and gentle northeastward structural plunge. The coarse to medium grained sandstone is interpreted to form the lower part of the Shickshock sequence.

The sandstone overlies the calcareous and silty slates. The contact is gradational and interbedded. The upper contact with metasedimentary and volcanic rocks is conformable. The latter are exposed along a lumber track between three to six miles east of Depot Jean, north of Matane River and along the northern shore of Lake Matapedia. In the latter occurrence the volcanic rocks occur within what appears to be synclinally folded arkosic sandstone. The sandstone makes up a band of variable width up to the Mount Albert area along the northern margin of the Shickshock Mountains.

Grain size of the sandstone is variable. Near Matane Lake, Mattinson (1958) found a three to twenty feet wide "augen" horizon in the sandstone. He considered it as due to deformation of a pebble bed in the sandstone. Generally, however, the grain size varies from coarse to medium.

A visual estimate of microscopic examination revealed an approximate percentage as follows:

Plagioclase	40 mostly authigenic, some rotated, An <sub>1</sub> -An <sub>10</sub> .
Quartz	20
Muscovite	10
Micaceous minerals	25-30
Epidote	4-5
Perthite, K-feldspar	2-5

The matrix of the sandstone is recrystallised and schistose.

Metasedimentary Rocks: These are greenish grey epidote-chlorite bearing

sandstone arkosic. Mattinson (1964) divided the metasediment into five zones, established in view of their having formed separately in temporary lulls in the volcanic activity which gave rise to Shickshock volcanic rocks. In his interpretation of the Shickshock structure, these zones are shown to be edgewise, resembling a section turned so it can be examined in plan.

The worker differs from this interpretation simply because no indication of steeply pitching lineation corresponding to an early phase of cross folding were observed in the Shickshock Mountain area. The axial lineations corresponding to the different major phase of folding are parallel, subparallel or oblique at small angles. The lenticular nature of the metasediments is solely due to the nature of its deposition in restricted basins.

A strongschistosity in the matrix is the most prominent structural fabric observed in the sandstone. It is marked by a pronounced recrystallisation of micas parallel to bedding, and authigenic plagioclase. The schistosity is considered tectonic. Quartz is often rotated and has frayed edges. Green biotite is common and resembles chlorite.

The age of the rocks is uncertain owing to the absence of paleontological evidence. McGerrigle (1954) regarded it as older than Middle Ordovician, i.e. Lower Ordovician or older solely because of style of deformation and stratigraphic distribution.

Ollerenshaw (1963) used micas in the arkoses for radiogenic age determination. The K-Ar method revealed a minimum age of  $530 \pm 35$  m.y. for the deformation which gave rise to recrystallisation of the micas. According to Kulp's scale (1961) it is Upper Cambrian.

Mattinson (1958), in view of the albitic nature of plagioclase, regarded the sandstone as derived from a landmass other than the more calcic gneisses of the north shore of the St. Lawrence. However, the plagioclase is mostly recrystallised and authigenic. In the chlorite-biotite association Na-rich plagioclase is common; calcic plagioclase, being unstable, breaks down into Ca-rich silicates. It is hence considered that a determination of the provenance of the sandstone from the plagioclase composition is not valid in view of the nature of the plagioclase.

The following is the visual estimate of percentages of the constituent minerals in thin sections studied:

Plagioclase	45-50
Quartz	15-20
Green biotite	2-5
Mica and sericite	5-10
Chlorite	5-7
Epidote	5-10
K-feldspar	2

Metavolcanic Rocks: The Shickshock metavolcanics are finely recrystallised lavas. These are bright to yellowish green in colour, and mainly consist of albite, amphibole, and epidote (Mattinson 1958).

Quartz is common, especially as intricately folded fine milky veins. The quartz veins are parallel to the mineral elongation and the schistosity. Authigenic pyrite is common. It grows with the long diagonal parallel to schistosity.

The volcanic rocks contain fragments of grey slate and also contain thin bands of highly sheared and fissile metamorphosed slate. Mattinson traced such a band for 1000 feet east of Mount Logan. A similar development of shale was observed in the metasediments along the Duvivier Stream.

The rocks are subophitic under microscope with laths of plagioclase usually with altered edges, and consist of approximately the following percentages (Mattinson 1958):

Blue-green amphibole	44
Plagioclase	22-26
Epidote	17
Chlorite	7-9
Quartz	.5-4

In the central and southern Shickshock Mountains the metamorphic rocks have a distinct bright green appearance. The rock is hard and attains a welded appearance as one approaches the southern margin. It is less fissile in the southern areas than the northern corresponding lithologies. A foot traverse along the western shore of Lake Matane is revealing in this respect. In the southern end of the traverse the rocks show flowage, consequently schistosity is not clear.

Nodules of yellowish green pillow-like masses with quartz vein distributed throughout were observed. These have cabbage-like concentric sheaths of foliation. The core is aphanitic and lacks foliation. This could be a metamorphic segregation feature (Mattinson, 1958) but the writer does not rule out the possibility of their being relict pillows. Rounded inclusions (2 mm. dia.) of epidote, quartz, chlorite, calcite, and albite are also found. They are prismatic in habit and occur in the higher metamorphic grade in the southern parts of the Shickshock Mountains.

Plagioclase in the mineral assemblage is usually replaced by amphibole which proceeds inwards from outer edges, but it is in turn replaced by chlorite which reduces the mineral into a turbid greenish mass.

Original volcanic layering is not seen in the Shickshock region. The layering observed is marked by a variation in colour from green to dark green or black, and probably is due to metamorphic segregation. The schistosity is parallel to layering in every observed exposure of the metavolcanic rocks.

In the vicinity of the Matapedia Lake, the Shickshock metavolcanic rocks are relatively unmetamorphosed, and are associated with arkosic sandstone and phyllitic grey slates. There is a predominance of arkose within the group. In the volcanic rock pillow structure are common, usually associated with jasper and calcite. Vesicles are abundant. Laths of plagioclase are often observed in the volcanic rocks.

AGE OF THE SHICKSHOCK GROUP: No fossils have been reported from the Shickshock group of rocks. The northern, probably lower contact is conformable with grey slates of Cambro-Ordovician age. (MacGregor, 1962). There are fragments and interbeds of slate in the arkosic sandstone of the Shickshock Group. The Shickshock metasediments and the metavolcanics are interbedded in the Matapedia Lake area and also in the western edge of the Shickshock Mountains near Depot Jean.

The southern contact with Silurian rocks is faulted.

K-Ar radiogenic age determination on muscovite in the schistose arkosic sandstone yielded a Middle-Late Cambrian age (Ollerenshaw, 1963). A similar age determination on the metamorphic aureole of the Mount Albert intrusive (50 miles to the east of the area) gave an age of  $495 \pm 35$  m.y. (MacGregor, 1962). The ultrabasic body apparently intruded already schistose Shickshock metavolcanic rocks that are older than Lower Ordovician.

Twelve miles southwest of Lake Matapedia, east of village La Redemption the Cambrian-Ordovician rocks occur as an inlier surrounded by the Silurian rocks on the north, and to the south, the Devonian rocks with a faulted contact. The rocks are schistose and phyllitic, with the foliation parallel to bedding. These are intruded by a serpentine-rich ultrabasic body which is elongated parallel to the schistosity of the host rock, and is itself cut across by a few thin dykes of pink granite (Beland 1960).

The ultrabasic body is probably equivalent in age to the Mount Albert intrusive, in view of its composition, and also owing to

its sill-like ultrabasic bodies lying outside the area to the east (MacGregor, 1962), in the Mount Sud area in the Shickshock Mountains. That it intrudes the phyllitic grey slate shows that the slate exposed near La Redemption is pre-Lower Ordovician in age.

CAMBRIAN-ORDOVICIAN SLATES: These are distinguished from the overlying sequence of rocks by a higher grade of metamorphism and a complex multiphase structural style. The rock lithologies somewhat similar to the overlying succession, are exposed in a belt between three to three and a half mile wide, immediately to the north of the Shickshock Mountains. These rocks outcrop southeast of the Shickshock Mountains up to Matapedia Lake. A linear SW-trending inlier of the rocks is exposed east of La Redemption.

The rocks comprise grey, black and greenish grey phyllitic sericitic slates with occasional fine calcareous or silty interbeds. The schistosity when refolded allows the rocks to split into tough but small pencil shaped cleavage fragments. The foliation surface is silvery in lustre. The bedding is often marked by laminae of siltstone with a welded appearance. The rocks often consist of interbedded shale and limestone. The thickness of limestone beds varies from thin lamination to a foot.

Limestone usually interbeds the slates in minor proportions. However, massive limestone although rare in this group of rocks, contain minor intercalations of slate. The slate in such occurrences is always schistose parallel to bedding. A massive limestone is exposed along the roadside section north of Matane River. It is steeply dipping and approximately 100 feet thick, white to grey in colour, fine grained, dense crystalline. It is interbedded with dark grey slate. In the same locality a grey oolitic limestone bed is deformed into series of boudins (Photo 3).

This belt of rocks is the only area where recumbent folds and horizontal or subhorizontal cleavage and schistosity is observed outside the Shickshock Mountains. The sub-horizontal fracture considered corresponding to the schistosity often cuts across the steeply dipping limestone beds (Photo 14). Multiple structural elements within the rocks are frequent recognisable on careful examination.

SCHISTOSE BRECCIA AND QUARTZITE: North of the Mount Albert, east of the study area an outcrop of schistose breccia was observed. It consists of fragments of limestone and sandstone of maximum length of one foot and an average length of five inches elongated parallel to bedding and schistosity which are parallel. The matrix is schistose and consists of slate and calcareous sandstone. The limestone fragments are fawn to grey in colour and micro-crystalline.

The conglomerate is associated with white quartzite, dense compact medium grained with faint cleavage fracture.

AGE OF THE CAMBRIAN-ORDOVICIAN SLATES: McGerrigle (1954) in Courcellette area, northeast of the study area observed a belt of complexly deformed rocks. Using sedimentary structures for top determinations he considered that the likelihood of the succession being inverted were just as much as that of its being a normal one. On structural relations alone McGerrigle considers it to be Lower Ordovician or older (p. 24).

A similar belt of complexly deformed rocks suspected to be inverted occur along the northern Shickshock Mountains and north of Matapedia Lake area. This belt lacks fossils; only a few fossils have been observed, and even fewer identified. Ollerenshaw (1963) found an incomplete cranidium of a trilobite identified as *Phychoparoid* in a section half a mile north of the Shickshock Mountains (fig. 6). These rocks are therefore, considered Middle to Upper Cambrian.

The Shickshock Group of rocks dip southeast. Hence, assuming the succession is normal, and the northern contact of the Shickshock rocks is conformable (as suggested by MacGregor 1962, Ollerenshaw 1963, and concluded by the worker) the rocks lying to the north of the Shickshock Mountains are older than the Shickshock Group. Accepting the radiogenic Middle-Late Cambrian age (Ollerenshaw 1963) for the Shickshock Group, the underlying occurring to the north should be of an older age, perhaps Early-Middle Cambrian.

However, there is sedimentary evidence that the beds lying to the north of the Shickshock Mountain area may be inverted. If so, the beds are younger than the Shickshock Group and are Late Cambrian-Early Ordovician in age (see structure of the Shickshock Mountains).

The contact of the Cambrian-Ordovician rocks with the younger

rocks is not clear. It is possible that it is marked by an extensive development of conglomerate or thick quartzite beds of Kamouraska facies or it consists of Impure Grey Sandstone Formation of the Quebec Group.

#### ORDOVICIAN ROCKS:

QUEBEC GROUP: The Quebec Group is named after the type section near Quebec City, and is essentially Middle and Lower Ordovician. The group consists of impure grey sandstone, grey to dark grey slate, often interbedded with silt and limestone, limestone, and quartzite conglomerate and associated limestone breccia of the Kamouraska facies. The name, previously used to comprise all slaty rocks of the area, excludes the Cambrian-Ordovician sequence of slate, sandstone, metavolcanics and metasediments. The Quebec Group is described formation wise in the following pages. The sequence of description does not necessarily follow the stratigraphic order as the relative ages are still open to criticism and question.

Impure Grey Sandstone: McGerrigle (1954) and Beland (1957) place the lower boundary of the Quebec Group at the base of an impure grey sandstone. The sandstones <sup>are</sup> locally gritty and gravelly and contain rounded shale fragments up to half an inch in diameter. The rock is usually medium to coarse grained, shaly, greenish-grey sandstone. When deficient in shaly constituents it is hard and compact. The mineral composition is variable. Following percentages indicate the range of composition in different localities: (Appendix II, Table 7):

Quartz	31.8-58.0
Feldspar	2.8--8.5
Calcite fragments	.9- 5.6
Volcanic fragments	.25-3.5
Chert	0-7.7
Matrix	27-61.6

Quartz grains are poor to well rounded. Sphericity is variable. Sorting is poor. A point count on the percentage of the volcanic fragments close to the upper contact faintly suggests an increase of the

volcanic fragments possibly lay to the south (See Appendix II).

The upper contact of the impure grey sandstone with the overlying grey slates is gradational. Thin layers of shale are frequent within the sandstone.

The sandstone formation is exposed in the Grosses Roches area, northeast of Matane in an anticlinal structure. The northern limb of the structure is mostly under water in the St. Lawrence, but three miles northeast of Ste. Felicite, the northern limb is exposed along the coast. It is inclined at 65°, overturned to the southeast (Photo 92). The overturned anticlinal structure is cross faulted at several localities. Another minor occurrence of the sandstone in the area is along the northwestern part of the area, northeast of Mont Joli along Point Metis.

The sandstone when rich in shale matrix is cleaved. The cleavage is of fracture type. It swings around the clastic grains. It is well jointed. Among the sedimentary structures it demonstrates graded and convolute bedding.

East of the study area a similar sandstone series known as Pillar Sandstone (Logan, 1846) are exposed in a belt lying to the north of Middle Ordovician rocks. The rocks are interpreted to young southward; hence, the sandstone is believed to underlie the Middle Ordovician rocks (McGerrigle, 1954). On the basis of structural evidence and the fact that upper contact with Middle Ordovician is gradational, the Pillar Sandstone are regarded as the basal bed of the Ordovician succession.

Graptolites in the shaly beds within the sandstone consist of Phyllograptus angustifolius Hall, Diplograptus dentatus (Brongniart), Didymograptus acutidens Lapworth, Climacograptus pungens Ruedemann, Trigonograptus ensiformis (Hall) (McGerrigle 1954). The sandstone on the basis of the graptolitic fauna are considered as equivalent to the Levis Group of formations of the Quebec area which belong to the Lower Ordovician, and suggest a "Diplograptus dentatus" zone (McGerrigle 1954, p. 30).

GREY SLATE: These comprise of grey to dark, black red and green slate,

calcareous slate interbedded with fine grey to fawn coloured argillaceous limestone, limestone or grey siltstone.

The Slates are variable in colour and composition. Along the coast, north of Mont Joli, they develop as hard and splintery pale green, grey calcareous shale and mudstone. These are openly folded and contain an axial plane cleavage. The cleavage is a fracture spaced at .2 to .4 inches. These grade into grey and dark grey slate containing interbeds of limestone. The slates contain a structure in which the cleavage fragments appear broken up into irregular pattern (Photos 65 & 66). On a strike joint surface individual fractures are restricted to within two cleavage fractures and do not continue into the adjacent fragment. They resemble a brickwork. The fractures are clean and never filled in by clastic material.

Grey slates laminated by calcareous beds often contain a secondary fibrous calcite vein along cleavage or bedding plane. These consist of calcite fibres at right angles to the surface of the vein. The vein in section is always divided in half at the middle. They are up to 1 inch thick but average thickness is approximately 0.5 inch.

The slates laterally and vertically grade into red and green variety. The red and green colour variation is due to secondary effects as the colours -

- (i) run discordantly to layering, and
- (ii) red colour often shows blotchy green colouration.

These are tough calcareous and also silty in composition. The cleavage is always widely spaced. The red colouration is due to hematite staining.

They develop at several localities along the St. Lawrence coast. The best development is approximately fifteen miles southeast of Matane. Other outcrops are four miles southeast of Baie de Sables; and extensively near Lake Portage and therefrom as a discontinuous mapable northeast trending band.

The slates when minor in proportion and interbedded with massive calcareous beds, contain a well developed bedding foliation.

Folds in the slates range from open to tight isoclinal, depending on lithology. The cleavage varies from coarse fracture in calcareous beds to preferentially oriented micas in non-calcareous beds.

The laminated slates and the slates interbedded with limestone and siltstone are usually well foliated. Most of the structural data has been collected from such rock types which contain similar anisotropy to allow the study of the deformed marker beds.

Brecciated and sheared zones of local significance within the slates are common especially at the contact between the slate and the massive beds.

The limestone and the siltstone interbeds are convolute bedded and cross bedded on small scale. The siltstone layers are often graded bedded.

Calcite veins are common. Quartz veins are rare. The calcite veins are usually parallel to the ac joints.

The grey slates overlying the impure grey sandstone lack fossils partly due to the intense development of cleavage, thus their age considerations are approximate.  $2\frac{1}{4}$  miles south of Matane graptolites, notably Climacograptus parvus Hall characterise the Dicellograptus zone (de la Rue, 1941).

Graptolites also indicate a Middle Ordovician age of grey slates exposed in the Shickshock Fault zone (see Appendix V).

**KAMOURASKA FACIES:** The rocks of the facies are named after their resemblance to the development of quartzite conglomerate near Kamouraska, L'islet and Montmagny counties southwest of the study area (de la Rue 1941, Ollerenshaw 1963). Dresser (1912, p. 14) proposed the term Kamouraska Formation.

The facies include calcareous sandstone, quartzite and associated limestone conglomerate in scattered lenses within the slate facies of the Quebec Group. These lenticular bodies of quartzite-conglomerate occur from Vermont, Southern Quebec to Newfoundland, and have aroused many a controversy.

The clastics occur in a repetitive manner within a preponderantly slate sequence and range in age from possibly Lower to Middle Ordovician.

Logan (1863, p. 265) reported the presence of conglomerate along the St. Lawrence. Near Metis clasts of up to twenty five tons

in weight are reported (Richardson 1859). Good outcrops occur near Grosses Roches in the area. These outcrops occur as lenticular and V-shaped bodies. Outcrops of the facies also occur near Goupil township east of the Riviere Matane village, and extend NE and SW therefrom. The outcrops vary in length from a few hundred feet to those up to five miles, and the width varies from less than a foot to 300 feet.

Over length of outcrops it shows little change except for local variation in fabric, and consists typically of sub-rounded to rounded lithified clasts of fine grained grey, weathering limestone, scattered phenoclasts often tabular of calcareous siltstone and sandstone, silty limestone or dolomitic limestone. In addition it also contains coarse crystalline dolomites. Clasts of cobble and boulder size are common. Matrix usually calcareous arenite surrounds the phenoclasts and makes up 30-35% of the total composition. The matrix is sometimes quartzose.

The quartzites grade into the conglomerates interbedding and lateral interfingering of the two is common. They are quartz cemented and rarely show coarse gritty to pebbly layers. They are white to grey, dense, and lack internal bedding structures. Average grain size is .12 mm., and grains are well rounded and well sorted and are tightly packed. Quartz makes up 90-95% of the clast, the remaining constituents consist of calcite, siderite, plagioclase magnetite.

The conglomerates vary from polymictic rounded conglomerates to intraformational breccia conglomerate containing oligomictic tabular fragments. The matrix varies from greenish shale to coarse calcareous sandstone. These lithologies intertongue. However, the interbedding between the Kamouraska and the slate facies is relatively rare. Sharp contacts are common.

In Goupil area the facies outcrop for about four to five miles in length and consist of boulder, cobble sized phenoclasts interbedded with quartzite and intraformational limestone conglomerate. Beds of slate are frequently sheared within this sequence and are extremely variable in thickness. Sedimentary structures are rare. The contact of the Kamouraska facies rocks with slate is usually extensively sheared.

Ollerenshaw (1963 p. 71) noted an occurrence of conglomerate at a slate-quartzite contact. The limestone conglomerate and the conglomerate show a development of axial fracture cleavage in shaly or calcareous sandstone matrix.

Folds within the fragments were observed by the worker but good examples were not found. Ollerenshaw (1963, p. 80) notes a 1-foot long phenoclast containing the hinge of an isoclinal fold. This evidence indicates presence of a tightly folded source area for these clastics. Judging from the size of the clastics the source area was not far.

A variety which is intraformational to subintraformational in composition contains flat to tabular grey calcarenite, silt, oolitic clasts, limestone, brown - weathering siltstone, microcrystalline grey dense limestone up to ten inches along the longest axis enclosed in green shale to fine-to-medium grained calcarenite matrix. It shows gradation from quartz sand to an oolitic type of matrix. The dense limestone fragments with brown siltstone are moulded and indicate a deposition in plastic state. These usually are between two to five feet thick.

- (i) The facies are deposited in shale and sandstone environment.
- (ii) Influx of Kamouraska facies is sudden.
- (iii) A conformable lower and upper contact with shale is usually disrupted and sheared during folding.
- (iv) Interbedding takes place between quartzite and limestone conglomerate.
- (v) Contemporaneous deformation produced minor conglomerate of intraformational aspect. These are excluded from the discussion of the more important aspect of the origin of quartzite and normal conglomerate-breccia.

There are several views regarding the origin of the Kamouraska facies within the slate facies:

- (1) That they are normal, well sorted, clean deposits formed on littoral marine beaches and in the shallow marine conditions, (Logan 1863, Dresser 1912, Hubert 1965) which were probably formed with rising fault scarp thus forming the coarse clastic debris (Alcock 1926, Blais 1950).

(2) That the Kamouraska facies are formed by a mixing of the shallow water coarse sediments with deep water facies marked by the absence of bedding structures, by turbidity current, instability of the basin slopes (Osborne 1934, Kindle and Whittington 1958, Bailey Collet and Field 1928). In addition to these effects it is also regarded as formed by sliding, slumping, penecontemporaneous movements or earthquake shocks (Baird 1960).

The following observations are relevant to any theory advanced to account for the origin:

(1) The outcrops of the quartzites do not occur at random within the slates. There is a linear discontinuous arrangement which probably marks a definite stratigraphic horizon.

(2) They are observed and investigated in the whole length of the Appalachian Range. The composition is remarkably uniform.

(3) A repetitive nature of the beds could be due to tight isoclinal folding in the areas. Since the quartzite lacks internal structure it is difficult to tell the tops of the beds. Hubert (1965) shows the presence of intricately folded quartzite beds of the Kamouraska facies.

(4) The associated intraformational conglomerates indicate local instability in the basin.

(5) Considering the size of the phenoclasts the source was local.

(6) Although the conformity between the underlying and the overlying slates is beyond doubt, the contact is often sheared and tectonic due to slipping during folding.

(7) They occur in lenticular bodies which could be sedimentary or tectonic in origin, and also probably the former modified by the latter process. These are often oblique to each other but do not join.

(8) The V-shapes of the outcrops are believed to be purely tectonic in view of:

(i) In the V-shaped outcrop the bedding of the underlying slates closely follow that of the quartzite.

(ii) The V-shaped outcrops often face each other. They are usually disconnected in the middle but appear as if

they were originally forged together but have been pushed apart.

(iii) In the outcrop pattern the hinge area of the V-shaped outcrops is thicker than the limbs.

(9) Absence of typical sedimentary structures rule out turbidity current origin.

(10) The immense size of the clasts, the scour structures indicate fast current regime in the fluvial transportation. This indicates large rivers and high relief. The landmass was probably freshly folded. This is also indicated by the presence of phenoclasts containing folds.

(11) The age of the facies is between Lower and Middle Ordovician (Ollerenshaw 1963).

In the present work the origin of the Kamouraska facies is considered as invoked by the first view listed earlier. These are thought to have formed as shallow marine littoral clastics in an oscillating shoreline; The source area was probably a folded uplifted landmass. The quartzite-conglomerate were formed as a reworked post-orogenic clastic deposits. These beds were tightly folded and consequently repeatedly exposed within the Quebec Group.

It is thought that it is possible that the quartzite-conglomerate sequence near Goupil, which is the most extensive of the developments in the area, and which also separates two distinct tectonic provinces, is significant to the tectonic history of the area.

Hubert (verbal communication 1967, and Ph.D. thesis 1965) believes that the clean sandstone of the Kamouraska facies signifies a marine shore environment. They mark a stratigraphic position and are continuous for long distances. According to him the presence of carbon in the matrix and non-orientation of quartz c-axes suggest that the quartz cement is secondary and has formed after the folding.

#### SILURIAN ROCKS:

Basal Conglomerate, Incognito Formation: Beland suggested a possibility of basal Silurian conglomerate on evidence of the presence of boulders containing subrounded to rounded fragments of the Shickshock volcanics

and the grey slate in a coral rich calcareous matrix. The very large boulders up to twenty feet across are found on the southern shore of Matapedia Lake near Val Brilliant village.

Ollerenshaw (1961) failed to find an insitu outcrop and called the possible bed incognito conglomerate. The worker favors the name as he also failed to encounter an insitu outcrop of the conglomerate in question.

Within the boulders the phenoclasts consist of blocks of arkosic sandstone, slate, and volcanic rocks which reach a maximum diameter of two feet. They are angular to subangular and elongated parallel to bedding. Sorting is poor. The volcanic fragments lack schistosity. The smaller fragments consist of the above mentioned rock type and also quartz and quartzite. The matrix makes up 20% of the rock composition and contains fossil coral bryozoa and brachiopods. Following fossils were reported by Ollerenshaw (1963):

Pentamerus sp.

Heliolites interstinctus

Favosite cf. F. niagarensis

Atrypa ' reticularis'

Clathrodictyon aff. C.

Cystosum

The fossils indicate the matrix to be of Upper Llandoveryan to Wenlockian age. Thus the conglomerates appear to be correlative to Awantjish- Val Brilliant interval or they may be correlated to Awantjish Formation (Ollerenshaw 1967) as local basal bed of the Silurian succession.

Awantjish Formation: Beland (1960) proposed the name for a succession of siltstone and sandstone, minor shale. The sandstone is grey to greenish with reddish spots on weathered surface. It is exposed southwest of Lake Matapedia directly overlying the Cambro-Ordovician slates, approximately two miles southwest of St. Cleophas. Although the lower contact is not exposed the sandstone and Cambro-Ordovician slates outcrop within half a mile. The former beds presumably unconformably overlie the latter.

The sandstone and siltstone are thick to massive bedded. The thickness is estimated to vary between 150-1000 feet.

Beland (1960) considered that the Awantjish Formation underlies the Val Brilliant quartzite, but the boundary is not clear as it is faulted against the St. Leon Formation. Ollerenshaw (1963) reported two occurrences of the Awantjish Formation nine miles NE of St. Tharsicius, along the Matane River. These consist of calcareous, medium to dark olive grey very fossiliferous shale with thin lenses of argillaceous limestone. It contains gastropods, crinoids trilobites and brachiopods. Presence of Resserella and Atrypa 'reticularis' indicates Upper Llandovery to Ludlow age (Boucot 1957). Monograptus cf. M. undulatus supports an Upper Llandovery age (Ollerenshaw 1967).

Val Brilliant Formation: Usually these quartzites form the basal beds of the Silurian sequence and lie directly on the grey slate sequence of Cambro-Ordovician age, especially in the vicinity of the Matapeda Lake area and between St. Tharsicius and the Shickshock Mountain edge (La Joie, 1961). In this area it attains a thickness of 300-500 feet.

In the area west of Tamagodi River the section of Lake Matapeda on the southern shore west of Val Brilliant village was used by Crickmay (1932) as the type section.

The quartzite is clean, white to fawn in colour, and typically red spotted. It ranges from coarse to medium in grain size, and contains occasional gravelly horizons. The quartzite is made up of 90% quartz. The loosely packed quartz grains are variable in rounding and range from well rounded to subrounded. Cement usually consists of quartz with up to 5% interstitial calcite. Hematite occurs peppered within the rock.

Towards the top of the formation the quartz cement is often minor to calcite. The rock is well bedded and ranges from fine to massive in thickness. The beds differ from each other in grain size, and on weathered surface by a reddish brown colouration. The lamination in the formation points to a rhythmic deposition. Fossil trails, grooves and bubble pits are common. The sandstones also show structures indicating penecontemporaneous deformation. These structures include load casts convolute bedding and piercement structures.

The Val Brilliant quartzite forms a synclinal structure southwest of Lake Matapeda, the northern limb of which is gently dipping at

between 10-20% to the south and the southern limb dips at up to 60°. Beland (1960) observed steep dips in the Val Brilliant quartzites in the vicinity of the serpentinite body southwest of Matapedia Lake.

North of the Shickshock Fault, Ollerenshaw (1961) observed several outliers of Val Brilliant quartzite along the strike of the Shickshock volcanic rocks. The dips are flatter (10-30°) than those of the underlying Cambrian-Ordovician slates which are tightly folded. An unconformable contact is suggested between the two rock types.

The Val Brilliant Formation contains following fossils (Ollerenshaw 1967):

Plicate Stricklandia of the gaspiensis type, Pentameroids and Porpites show Upper Llandoveryian to Wenlockian age. Local presence of Stricklandia triplesiana indicates a restricted, probably Upper Llandoveryian age for part of the Val Brilliant Formation.

Syabec Formation: It overlies the Val Brilliant and is restricted to the central part of the area. It is exposed along the Tamagodi River section northeast of Matapedia River and Amqui township.

It consists of a grey highly fossiliferous silty or clayey limestone with thin intercalations of silt or shale with wavy beds an inch or more thick. It is fine grained, and coarse crystalline calcite veins or fossils stand out in relief as resistant constituents.

The fossils consist of corals, brachiopods, gastropods and crinoids. The limestone is made up of small fragments of coral fragments and algae set in fine grained calcareous matrix. The limestone beds are mud cracked (La Joie 1961).

The Upper and Lower Contacts are not exposed but it is considered at least 500 feet thick by Ollerenshaw (1963).

Following fossils are identified (Ollerenshaw 1967):

Conchidium sp.

Leptaena rhomboidalis

Plectodonta ? sp.

Favosite sp.

Syringopora sp.

Presence of Conchidium indicates a Llandoveryian age. Thus stratigraphically conformable relationship with the underlying Val Brilliant

Formation suggest that the Syabec Formation extends from Wenlockian to Lower Ludlovian age.

St. Leon Formation: It is extremely well developed in the northern part of the Siluro-Devonian basin. The type locality is at St. Leon Le Grand on Humqui River (Crickmay, 1932). The formation comprises of mostly green, light to dark grey, red calcareous siltstone and sandstone along with green, dark grey and rarely red shale. It also contains beds of limestone and thin layers of conglomerate. The siltstone is laminated in places but generally it occurs in interbeds of massive beds intercalated by minor dark grey to black shale. The shale is frequently silty and calcareous with conspicuous muscovite. The limestone occurs in the siltstone as beds, a few inches thick, and is generally fossiliferous. It also occurs as discontinuous lenses along bedding.

A siltstone conglomerate occurs along the Amqui anticline near Amqui. Near St. Leon, limestone conglomerate is up to 100 feet thick. The conglomerate beds on both flanks are made up of rounded fragments of limestone, silty limestone and organic material. Most of the organic material, coarse and fine grained, consists of corals, algae, and crinoids.

The St. Leon Formation south of the Shickshock Mountains forms a continuous outcrop which ranges from two miles wide, northeast of St. Tharsicius to one-half mile wide near Truite brook. It is faulted against the Cambro-Ordovician slates and gives rise to a fault breccia exposed along the Matane River south of the Shickshock Mountains.

South of the Shickshock Mountains, four miles east of Matane Lake, the St. Leon Formation consists of calcareous siltstone beds of up to one foot thick intercalated with minor arenaceous limestone, with thin beds of shale and siltstone. This imparts the rock a slabby fracture. The rock in roadside exposure along Bonjour Brook, in the vicinity of the Shickshock Fault is folded into open monoclines, hump shaped folds (Photo 52). The rocks dip at 10°SE.

The sandstone and siltstone are frequently laminated, slumped and cross bedded in small scale.

Other sedimentary structures include ripple marks, convolute bedding, pseudo-conglomerate. Near the occurrence of pseudo-conglomerate

the bedding in limestone is deformed into knotted or boudin-like masses.

Along the Tamagodi River, Ollerenshaw (1961) estimated the St. Leon Formation to be 5000 feet. Beland (1960) estimated a thickness of 7,300 feet along Mestigouche River (lying to the southwest of the area). Stearn considered from the Associated Developments Causapschal oil well that the St. Leon Formation is at least 5000 feet thick.

In the southern part of the area the Silurian-Devonian is a complex assemblage of volcanic and sedimentary rocks which are described along with the Devonian rocks.

St. Leon Formation is not richly fossiliferous. It contains graptolites in the shale member, and coarser clastics contain brachiopod, coral, crinoid and trilobite.

The St. Leon Formation contains following graptolites (Ollerenshaw 1967):

Monograptus cf. M. colonus

Monograptus cf. M. tumescens

Monograptus ultimus

Monograptus cf. M. leintwrdinensis

Linograptus posthumus posthumus

The graptolite assemblage is of Ludlow age. The last mentioned fossil is of Eurasian species and occurs in Middle and more particularly, the Upper Ludlow. Stearn (1965) reported Lower Ludlow graptolites in upper part of the St. Leon Formation in the Causapschal area.

#### DEVONIAN ROCKS:

The Devonian formations, Cape Bon Ami and Grande Greve have often been mapped as a single unit. Crickmay (1932) included the two in Causapschal Formation and Beland (1960) in the area west of Matapedia River mapped the Devonian rocks as "Cape Bon Ami - Grande Greve limestone". Stearn (1959) and Ollerenshaw (1963) separated the two as distinct mapable units.

The presence of graptolites in the lower-most Cape Bon Ami member (Cumming, 1959) indicates that it is partly Silurian in age. The contact between the Silurian and Devonian rocks therefore lies within the Cape Bon Ami Formation.

Cape Bon Ami Formation: The gradational contact between St. Leon and the Cape Bon Ami Formation is poorly exposed along Truite River, half a mile upstream from the Matane River confluence.

The Cape Bon Ami comprises of shaly to silty and sandy limestone of grey to dark grey colour. Usually it is laminated showing compositional layering and a colour variation within the shades of grey. It is often massive bedded.

The limestone and silty limestone are rich in organic matter and contain a bituminous smell on freshly broken surface. The rock is fossiliferous and the fossil stand out in relief on weathered surface.

Cape Bon Ami Formation is arbitrarily separated from the overlying Grande Greve Formation on the basis of latter being a calcareous siltstone rather than silty limestone. South of the Shickshock Mountains the distinction consists only of a more silty nature of the Grande Greve than the Cape Bon Ami Formation.

The thickness of the Cape Bon Ami is difficult to estimate in view of its hinges being thicker than the limb areas due tectonic processes (Stearn 1959). An average thickness of 2,500 feet is computed for the Causapsal Lake area. Northeast of Causapsal town the formation dips vertically, and is 1000 feet.

Cape Bon Ami Formation contains following fossils (Collected by Ollerenshaw 1963):

<u>Coelospira virginia</u>	<u>Schuchertella</u> sp.
<u>Leptocoelia</u> sp.	<u>Chonetes</u> sp.
<u>Spinoplasia</u> sp.	

The fauna is Lower Devonian and Leptocoelia, Plectodonta, and Spinoplasia gaspensis indicates a New Scotland age (Ollerenshaw 1967).

The occurrence of Monograptus aequabilis in Lower part of Cape Bon Ami Formation suggests that the base of the Cape Bon Ami Formation lies near the Silurian-Devonian boundary (Boucot et al. 1967).

Grade Greve Formation: The contact between the Cape Bon Ami and the Grande Greve is gradational. The Grande Greve Formation consists of grey to dark grey hard flinty calcareous siltstone or fine grained sandstone, often slightly quartzitic. It has a rare conchoidal fracture.

The lower part of the formation contains closely spaced laminae of occasionally yellowish brown siltstone which have been etched out into a relief by weathering.

Bedding is difficult to trace in the upper part of the formation. Cleavage develops as irregular fracture and resembles a closely spaced joint set up to two inches apart, a good example which is along the road north of St. Alexandre des Lacs, east of Amqui.

The Grande Greve Formation is 3,500 feet thick, and the thickness is reduced to 2000 feet on the steep limb of the Causapschal anticline due to tectonic thinning.

The upper contact of the Grande Greve Formation with the overlying York River-Heppel Formations in north-central and eastern parts of the area is placed at the appearance of the first thick sandstone bed rather arbitrarily (Stearn 1959). McGerrigle (1954, p. 43), in Courcellette area east of the study area, introduces York Lake facies for the transitional beds between the Grande Greve and York River. These interbeds of clastic and calcareous rocks grade laterally and vertically into the York River Formation

In the area south of the Shickshock Mountains these interbeds are probably several hundred feet thick and consist of light coloured lenses of fine sand with an increased fossil content.

The fossils in the Grande Greve Formation are rare and poor in quality. McGerrigle (1954) considered that poor fossil record does not allow the determination of the relative position and zoning of the Cape Bon Ami and Grande Greve Formations.

The age of the Upper part of the Grande Greve Formation is Lower Devonian (Boucot et al. 1967).

**FORTIN SLATE GROUP:** Fortin Group of slates extensively developed as a central belt 16-18 miles wide consists of grey to dark grey slates, sericitic and often phyllitic. These contain calcareous and silty interbeds. Rarely it also contains greywacke up to several tens of feet thick.

In its typical development the rock is laminated. The layering is marked by a colour variation from grey to dark grey, or by a differentiation of weathering which results in opening of the fracture within the finer or non-calcareous layers (Photo 60). The

beds are often calcareous or massive in appearance. The sandstone beds are grey, medium to coarse grained, locally thin bedded showing gradation of grain size. Some beds are feldspathic. Grits, pebble conglomerate are found in isolated beds which contain fragments of limestone, quartz, slate, black chert and also fine grained sandstone.

Sedimentary structures consist of cross bedding in small scale, convolute bedding and slump bedding.

Sericite is authigenic and is present with chlorite. The sericitization and the faint chloritisation are the only mineralogical signs of metamorphism.

Fragments of bryozoa (?) fossils in recrystallised ground-mass were observed in rock thin sections. In general fossil content of the slates is poor, because of extensive development of cleavage and recrystallisation.

York River-Heppel Formations: The Grande Greve siltstone and sandstone grade upward into an interbedding of sandstone and siltstone which is termed as York River around Casault Lake and Heppel Formation near Heppel village (near Ste. Florence). It consists of an interbedded sequence of grey siltstone and sandstone at the top and bottom, each between 4,000 and 6,000 feet thick, separated by about 4,000 feet of thickly bedded reddish coarse sandstone. The red sandstone are similar to those described by McGerrigle (1954) as the middle member of the Gaspé Sandstone Formation in the north central Gaspé. The Heppel Formation of the central part of the study area contains the three units, York River, Lake Branch, and Battery Point divisions of the Gaspé Sandstone Series (McGerrigle 1954).

The York River-Heppel Formations are well exposed along the Causapsal River. The river cuts an up dip section, and exposes gentle dipping bedding surfaces at  $10^{\circ}$ , providing good fossil localities.

Marine fossils are particularly rich in a calcareous zone approximately 2,000 feet above the base. This zone contains brachiopods, pelecypods, bryozoa, trilobites, gastropods, and coral fossils of biohermal nature. Kindle (1938) regards it as late Devonian in age. Cooper (1942) considers it as partly Middle Devonian (Onandaga). Stearn (1959) assigned Middle Devonian age to the York River-Heppel Formations. Ollerenshaw (1963) showed it to be upper part of Lower Devonian in age.

A re-study of the base of the formations indicates that at least the lower 2000 feet of the sandstone is Lower Devonian in age. The brachiopod genera Rhenorenselaria, Prionothyris, and Globithyris are present in the sandstone, known elsewhere from the upper part of the Lower Devonian (Boucot et al. 1967).

Along the Matapedia Valley beds of the York River-Heppel Formations dip steeply to the south in an outcrop about three and one-half miles wide. The attitude of beds is irregular.

Beland (1960) attributed approximately 14,000 feet thickness to the York River-Heppel Formations. It is possible according to him, that the beds are progressively younger westward.

Ste. Marguerite Volcanics: A belt of igneous rocks is exposed close to the northern edge of the Fortin Slate belt, northeast of Ste. Florence village. These dark-green, fine-grained, amygdaloidal rocks were identified by Alcock (1935) as augite andesite. The igneous rocks were traversed along Fraser brook which starts between Ste. Marguerite and Ste. Florence villages and joins the Matapedia River, downstream from Ste. Florence. The igneous rock outcrop has been mapped as cut off by a cross fault (Stearn 1959) on its eastern margin. The western margin is in contact with the Ste. Florence fault.

Northeast of Ste. Florence along the Highway 6, an exposure of volcanic breccias lies in the strike of the outcrop of the volcanics. The breccia is composed of fragments of  $\frac{1}{2}$ -inch diameter of grey aphanitic rock in chloritic matrix, interbedded with chloritic slates. The chloritic slates are also found interbedded with the volcanics (Stearn 1965). The Ste. Marguerite igneous rocks are considered volcanic because (i) they are interbedded with foliated tuffaceous beds; (ii) they are amygdaloidal and brecciated, and (iii) appear to be genetically related to the volcanic breccia exposed near Ste. Florence.

The stratigraphic position within the Fortin Slate is not clear. The thickness is estimated at approximately 600-700 feet. The volcanics are well jointed.

## UPPER ORDOVICIAN - DEVONIAN ROCKS OF THE SOUTHERN PART OF THE AREA:

The Ordovician - Silurian - Devonian succession in the southern part of the area consists largely of fine clastics and volcanics. These have been mapped as Matapedia Group (Crickmay, 1932; followed by Alcock, 1935). The complex sequence of the sediments and volcanic rocks was thought to represent Upper Ordovician sequence as dated by trilobite and brachiopod fossil content found near the mouth of the Matapedia River. Although the rocks containing the fossil localities were truly Ordovician in age, the whole sequence was found to comprise of Ordovician, Silurian, and Devonian sediments (Beland, 1958).

The rocks consist of a belt of Ordovician rocks three miles wide at the narrowest part, widening to seven miles at the widest part along the Matapedia Valley. The increased width is presumably owing to a southwesterly plunging synclinal structure. The Ordovician belt of sediments is confined on the two sides by shale, calcareous siltstone, sandstone, intraformational conglomerate, algal or coralline limestone and dark grey shale siltstone, sandstone with minor grits of Silurian - Devonian age.

UPPER ORDOVICIAN MATAPEDIA GROUP: It is exposed in a northeasterly trending belt which consists of light to dark coloured silty and shaly limestone. The individual beds are of variable thickness. These contain shales with reddish brown silty intercalations. The calcareous shales often contain beds of shaly limestone with a rough conchoidal fracture. Bedding is often wavy in appearance and hard slabs break along cleavage parting. It is typically dark grey and dense in appearance. Calcite veins are common with occasional quartz.

Cleavage is steep and northwesterly dipping.

Along the Highway 6, near the bridge across the Matapedia River at St. Alexis, calcareous slates are exposed. These contain extensive calcite veins. These rocks are folded into a series of open to tight folds. Sills and dykes are frequent along the southern contact with the Silurian - Devonian rocks.

The fossils collected from Matapedia Group include the following (Alcock 1935):

Cyclopyge sp.

Cyclospira sp.

Winnepegoceras n. sp.

Atrypa sp. cf. A. marginalis

Illaenus sp.

Calymene sp.

The fossils are broken and imperfectly preserved, and include trilobite fragments, crinoid columns, bryozoans. The fauna suggests a late Ordovician age for the group (Alcock 1935).

Silurian and Devonian Rocks: To facilitate the description, the Silurian and the Devonian rocks can be distributed into those; (1) exposed to the north, and (2) south of the Ordovician belt of rocks.

The southern belt: A four-to five-mile wide band extends from southern to the eastern edge of the area. The rocks can be sub-divided into two belts lying: 1. Northwest of a thick volcanic wedge, grey to dark grey limestone often shaly and fossiliferous. 2. South of the volcanic wedge, the rocks are brown weathering, grey, greenish grey, fossiliferous calcareous siltstone, and slightly calcareous massive grey or green mudstone. The two rock types are interbedded. Coquina-like beds a few inches thick and coralline and algal rocks were observed southwest of St. Fidele.

South and southwest of the volcanic wedge several conglomerate lenses intercalate shale siltstone resembling the Fortin slates. A series of conglomerate beds of an irregular thickness of up to several hundred feet occur near the western tip of the volcanic wedge. These consist of rounded clasts of limestone with hard, fine grained sandstone. The colour is reddish, the matrix is shaly sandstone.

The volcanic zone is two to three miles wide. It consists of thick flows of basic, intermediate and rarely acidic volcanic rocks with intercalated tuff and agglomerate. They are mostly black, dark grey, green and aphanitic in texture and lack pillows. Amygdules of carbonate, black chalcedony, chlorite and zeolite are common in porphyritic or non-porphyritic types. Occurrences of acidic porphyritic rock with pink feldspar are rare.

North of the Ordovician belt the rocks resemble the Fortin sequence described earlier.

The rocks from the eastern margin of the area to l'Alverne township are mostly soft shales and fine grained quartz sandstone, calcareous shale, limestone and shaly limestone. Southwest of l'Alverne, the rocks are calcareous, fine grained sandstone, thin bedded limestone and rusty shaly siltstone. The rocks are well cleaved.

In proximity of the contact with the Ordovician rocks, the fine sandstone, well bedded grey, green, weathering brown lack cleavage. These are considered the transition beds of Silurian age which laterally and vertically pass into calcareous shale facies of the Fortin Group (Beland 1958).

**FORTIN GROUP** These are described by Beland (1958) in the Oak Bay area and consist of grey, medium to coarse, feldspathic sandstone with gritty and conglomeratic interbeds. These are accompanied with calcareous shale, thin bedded and occasionally interbedded with siltstone. These are tightly folded and the cleavage is well developed.

Lower or Middle Devonian Conglomerate and Sandstone of the Coastal

Area: These develop near Cross Point and the Campbellton Bridge. Pebbles in conglomerate include various types of volcanic rocks, tuff, black chert, calcareous siltstone and limestone. The clasts are fairly well sorted. The sandstones are green to grey, coarse grained, cross bedded and contain fossil plants.

They are Middle Devonian in age (Alcock 1935). McGerrigle (1953) reclassifies them as Lower or Middle Devonian.

**Intrusive Rocks:** Throughout the southern area the intrusive rocks consist of dykes and sills of intermediate to acidic composition. The intrusives are frequently close to the volcanic rocks. The thickness is variable. Porphyritic and amygdaloidal nature is common. The acidic intrusive bodies are relatively less common.

## APPENDIX II

## PETROLOGY OF THE IMPURE GREY SANDSTONE.

A petrological study of the impure grey sandstone of Lower Ordovician age was carried out mainly to study the percentage of the volcanic fragments and suggest a source for the same.

The rock consists of unsorted clastic constituents, the remaining being cement composed of calcite, clays with sericite and chlorite. Rounding and sphericity are poor. The average grain size is variable and beds of grits or pebbles are common. Generally, the average grain size dia. is close to .5 mm. Fragments of shale or pyritiferous shale are uncommon.

The rock is made up of the following constituents:

Quartz: It varies in rounding and sphericity and ranges from angular to subrounded. Sorting of the quartz grains is poor. Gneiss fragments are present. The grain size is also variable.

Chert: It is usually primary sedimentary, angular and unsorted.

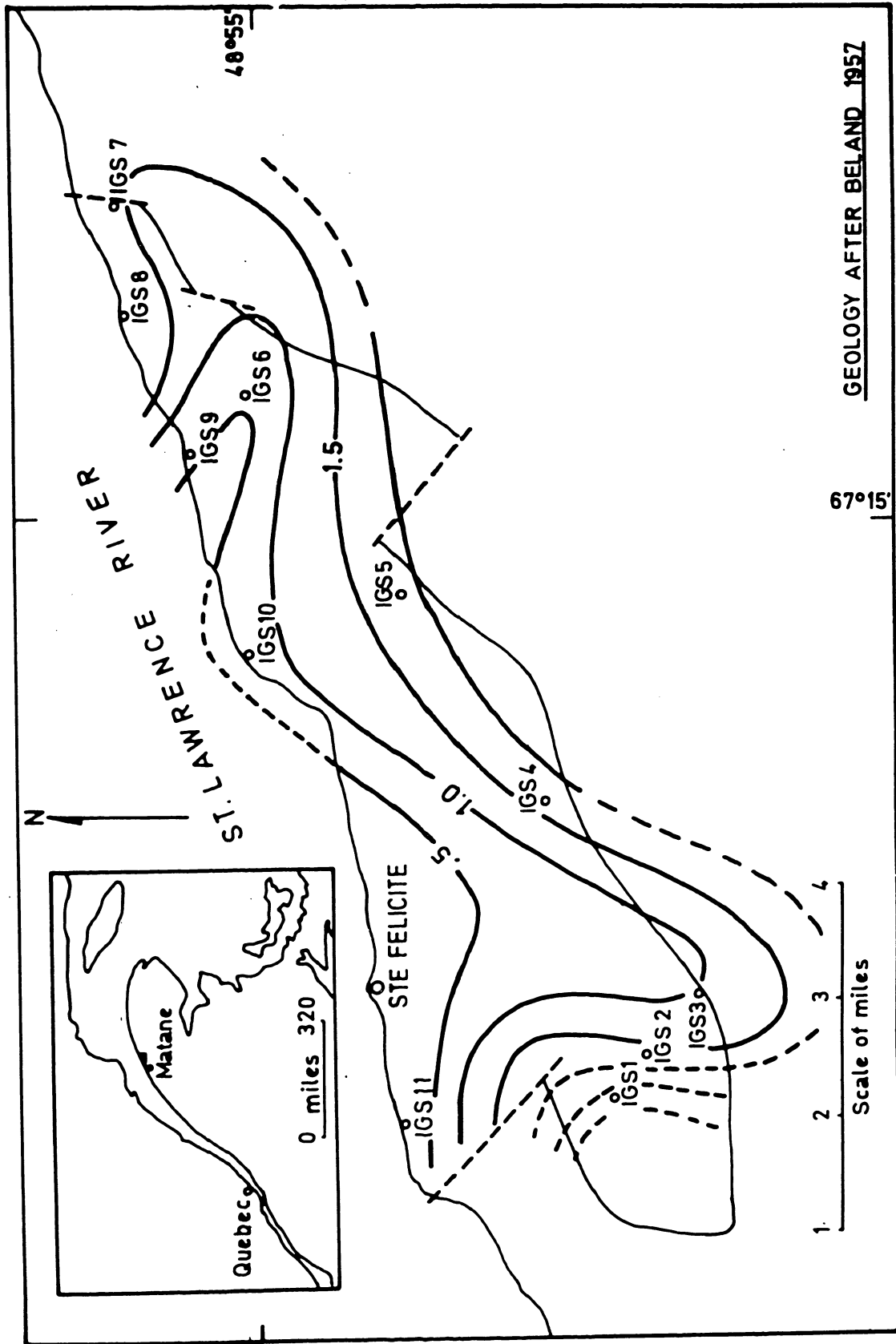
Limestone Fragments: The limestone fragments are usually large in diameter, and usually range between .5 and 1 mm. dia. They are angular. The fragments often show relict bedding and usually contain silt sized quartz grains peppered throughout the body. Sometimes they are argillaceous.

Igneous rock fragments: These are basic in composition, and are chloritised. The chloritisation is most intense on the outer margins of the grains than the centre. The internal structures in the basic volcanic rock fragments (R.I. > C.B.), especially the presence of the  $S_1$  schistosity if the rock has been derived from the Shickshock Group of the Shickshock Mountains, is difficult to determine, but it is considered absent in the few fresh grains that were observed.\*

Plagioclase: The mineral grains, generally .3 mm. in diameter, are usually

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\* Laths of plagioclase with altered margins demonstrate relict ophitic structure. Unaltered patches are present within altered grains.



**FIGURE 28. PERCENTAGE OF VOLCANIC FRAGMENTS IN IMPURE GREY SANDSTONE CONTOURED IN STE.FELICITE GROSSES ROCHES AREA.THE FRAGMENTS ARE PARTLY OR WHOLLY CHLORITISED.**

altered and chloritised. It is impossible to determine the optical properties of the plagioclase due to the intense alteration. The plagioclase is usually replaced by secondary calcite.

Detrital matrix: This constitutes a high percentage of the total composition. It has a green to brownish colour, faint to non-pleochroic. The chlorite in the matrix is green to yellowish green, with a low-order grey interference colour. Glauconite and micas are present but the percentage is low. The matrix sometimes contains fine-grained quartz grains. No schistose structure corresponding to the  $S_1$  schistosity in the Shickshock Group is observed in the rocks (microphotos 20 and 21).

The composition of the 11 samples of the Impure Grey Sandstone is given in table 7.

Table 7. Percentage Composition of the Impure Grey Sandstone for Samples from Different Localities:

	SGW 1	SGW 2	SGW 3	SGW 4	SGW 5	SGW 6	SGW 7	SGW 8	SGW 9	SGW 10	SGW11
Quartz	58	48.5	42.3	51.5	35.6	31.8	46.0	38.8	42.0	61.2	39.5
Feldspar	2.8	5.0	3.8	4.5	5.4	4.6	8.5	4.2	8.6	3.5	2.9
Calcite frags.	5.6	4.4	4.3	1.9	4.3	1.3	1.2	6.3	0.9	4.9	2.2
Volcanic frags.	3.5	1.5	0.9	1.9	1.8	0.5	1.5	6.0	0.6	0.8	0.2
Chert	3.1	3.7	1.8	0.9	1.2	0.0	0.8	7.7	0.8	1.2	1.7
Matrix	27.0	36.7	47.0	39.3	51.6	61.6	42.0	37.0	47.0	28.2	53.3

CONCLUSIONS: It is concluded from the petrological study of the impure grey sandstone that:

1. The basic volcanic rock fragments in general were probably not derived from the Shickshock Group of the Shickshock Mountain area, because they lack the  $S_1$  schistosity found in the Shickshock group of rocks.

2. The igneous rocks probably were derived from the numerous but minor volcanic intrusions within the Ordovician rocks.
3. The provenance of the igneous fragments probably lay to the south as suggested by the gradual increase in the fragment percentage to the south (fig. 28).
4. The low percentage of the igneous rock suggest a great distance between the source area and the outcrop area examined.
5. Unsampled gritty and gravelly beds with rounded to sub-rounded phenoclasts in the formation suggest occasional influx of coarse material.
6. The absence of any schistose structure from the matrix or the sandstone itself suggests that it has not undergone the intense deformation indicated by the structures in the Cambrian-Ordovician rocks, and the Shickshock Group.

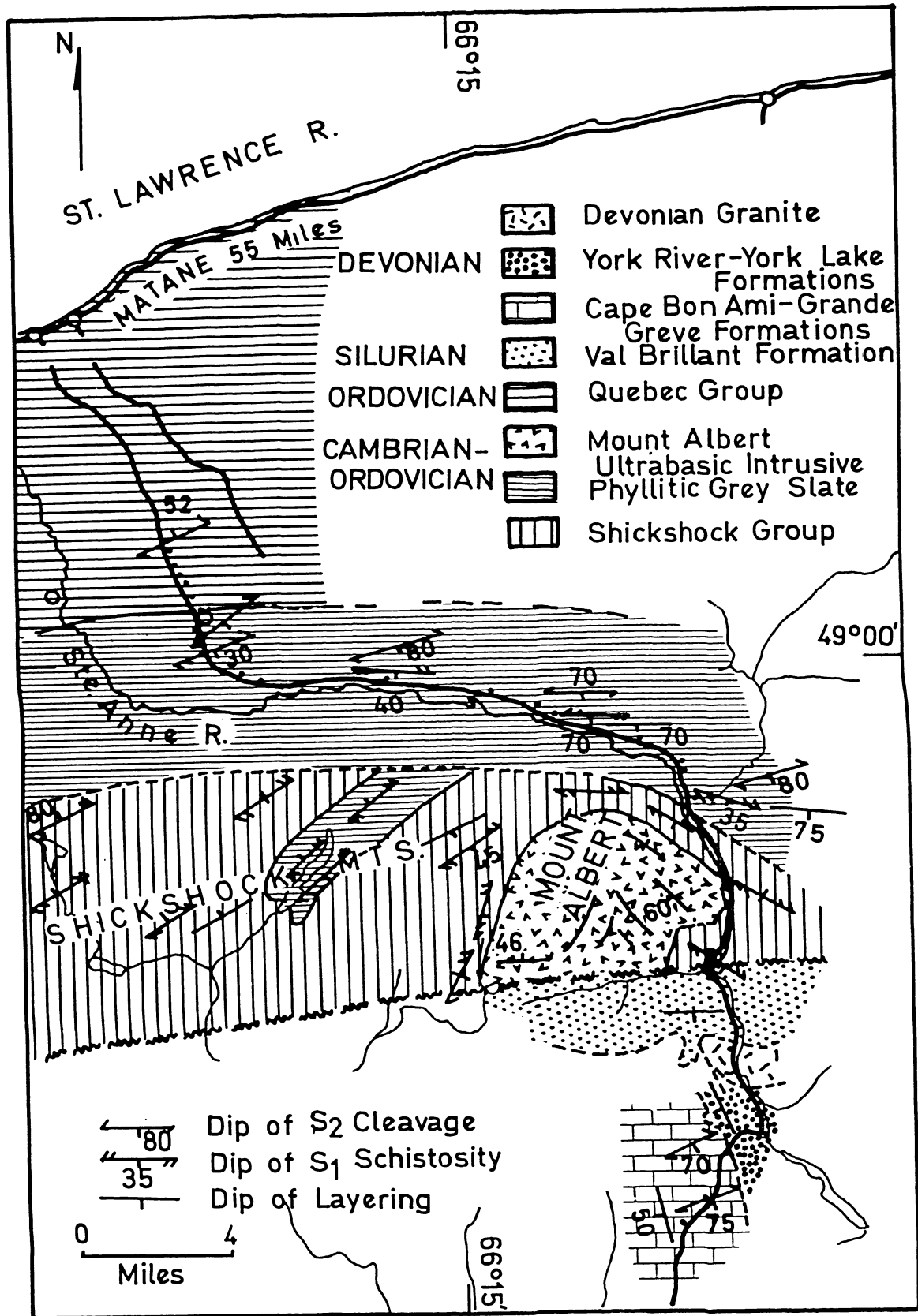


FIGURE 29. GEOLOGICAL MAP OF AREA AROUND MOUNT ALBERT (AFTER MacGREGOR 1962, AND ROBERT 1965)

APPENDIX III.

## THE TRAVERSES NORTH AND NORTHEAST OF MOUNT ALBERT

Mount Albert is situated approximately 63 miles east of Matane Township. From the coast, it is accessible from the Ste. Anne des Monts township by the Trans Gaspe Highway (fig. 3).

The attention to the area was drawn by the geological map of the area near Mount Albert by McGerrigle (1954) and MacGregor (1962). According to the latter worker the  $S_1$  schistosity is older than the Mount Albert ultrabasic intrusion. The  $F_2$  folds that deform the  $S_1$  schistosity, skirt around the igneous body. The swing of the  $F_2$  trend is in a gentle curve, and it maintains a general ENE trend (fig. 29).

The northern boundary of the Shickshock Group which in the area investigated in the present study trends NE and ENE, bends E-W, and then NW-SE. The exposure finally lenses out. This is anomalous to the  $F_2$  trend, therefore it was suspected that probably the earlier fold system, present in the Cambrian-Ordovician rocks was cross folded. This required a closer examination, and the field work was carried out in the summer of 1966.

The traverses were made mostly along the Trans Gaspe Highway, because it offered some of the best exposures in the area. The Highway follows the Ste. Anne River, which south of the Shickshock Mountains, flows around the eastern edge of the Mount Albert summit of the Shickshock Mountains, westward, north of the Mountains, then, finally, northward before joining the St. Lawrence River.

Continuing a traverse from Ste. Anne des Monts, the  $F_1$  and the later,  $F_2$  folds are parallel, until a point north of Mount Albert is reached. East of this point the  $F_1$  structures diverge as more easterly trending than the ENE trending  $F_2$  structures.

Following is the description of the outcrops in the traverses from SE to NW:

St.A-1. The  $F_1$  structures are best exposed under the bridge on Ste. Anne Northeast stream under the old section of the highway. The

rocks consist of grey to greenish grey, phyllitic slate with a silvery lustre. Here the  $S_1$  schistosity trends at  $120^\circ$ , dipping at 30-40 SW. The  $S_2$  cleavage cuts across the  $S_1$  obliquely, dips steeply SE; and the  $L_2(S_1/S_2)$  plunges westward (photos 9 and 10).

Further south of the confluence of the Ste. Anne and the Ste. Anne Northeast Rivers the Shickshock Group and the Associated grey slates consistently trend SE.

St.A-2/1--2/3. These localities are situated 200 yards south of La Gite, the hotel of the Gaspé National Park. In this locality an interbedded contact between the grey phyllitic slates and the Shickshock meta-volcanics is exposed. The schistose metavolcanics contain the  $S_1$  schistosity, and are interbedded with grey and red slates. North of the contact are the slates and to the south, the metavolcanics. On the  $S_1$  surface of the metavolcanics dark green lensoids of chlorite are found. A fine striation trends parallel to the long axis of the chlorite lensoids. These are considered parallel to the  $F_1$  axes. The interbedded slates have a baked appearance. An irregular  $L_4$  develops on the  $S_1$  surface.

St.A.-2/4. In this outcrop phyllitic grey slate with a silvery lustre is exposed. The  $S_1$  is parallel to the calcite layers, which are re-folded by the E-W trending  $F_2$  folds. The  $S_2$  is poorly developed.

St.A.-2/5. The  $S_1$  structures trend N80E, and are suspected as parallel to the  $F_2$  structures.

St.A-2/6. The slates are highly plicated, and tightly folded. The  $F_2$  folds deform the  $S_s$  and also the  $S_1$  schistosity. The  $S_2$  cleavage is coarse and steep. The  $S_s$  and  $S_1$  trends N60E and dips vary from 75SE to 65NW.

St.A-2/7, 2/8. It is a small roadside exposure, in which the  $S_1$  is well developed and  $L_2$  pitches SW on it.  $S_s$  and  $S_1$  are clearly oblique.

2/8 is exposed NW of 2/7, and consists of schistose conglomerate. The phenoclasts are composed of limestone and are flattened parallel to the  $S_1$  schistosity. The matrix consists of calcareous slate, and is schistose (Photo 13). This is regarded as  $S_1$ , a schistosity corresponding to that in the Shickshock metasediments. It is steeply dipping and parallel to bedding.

St.A-2/9, 2/10. The exposure consists of grey slate with limestone interbeds. The  $S_s$  and  $S_1$  are parallel and refolded by the  $F_2$  fold.

St.A-2/11. It is a large roadside exposure and a subhorizontal fracture varies in inclination from a low angled southeasterly to northwesterly dip. In the limestone beds, which usually are devoid of a well developed fracture, a coarse and consistent fracture corresponding to the  $S_1$  is observed (Photo 14). The  $S_1$  is refolded by the upright  $F_2$  folds. A steep southeasterly dipping  $S_2$  fracture is present.

St.A-2/12. This outcrop consists of grey phyllitic slate. It is schistose. The  $S_1$  and  $S_2$  are probably parallel. The  $L_1$  and the  $L_2$  consist of a trace of bedding on  $S_1$ , and a crenulation lineation respectively.

St.A-2/13. The rock consists of calcareous grey slate. A subhorizontal fracture regarded as corresponding to the  $S_1$ , cuts across the southeasterly dipping  $S_s$ . The  $S_2$  is upright and poorly developed (photos 8 & 8a).

St.A-2/15 The rock consists of grey slate with calcareous beds. The folds are tight isoclinal, reclined, and are considered  $F_1$ . The axial planes are low dipping (photo 1). These are refolded by a steep and widely spaced  $F_2$  folds. The  $S_1$  is parallel to the  $S_s$  except in the hinge areas. The  $S_2$  is poorly developed.

St.A-2/16, 2/17. In this outcrop of grey slates, the  $S_s$  dips at NW and SE, and the  $F_2$  folds trend N50E. The  $S_1$  is probably parallel to  $S_s$ .

St.A-2/18. It consists of massive calcareous beds in slate. The slate has a prominent bedding foliation.  $S_1$  is not clearly observed, and  $S_2$  is poorly developed.

St.A-2/19. The exposure consists of grey phyllitic slate with limestone intercalations. The metamorphic grade is similar to that of the grey phyllitic slates exposed in localities already described. The structure is complex, presumably contains a folded  $S_1$  schistosity, but it is not clearly indicated. The outcrops north of 2/19 are earthy in appearance and have a simpler fold style than those exposed in the south.

St.A-2/20. This outcrop consists of limestone breccia of Kamouraska facies. The matrix is calcareous slate and fine arenaceous fractions.

The phenoclasts are randomly oriented, and are angular and rounded fragments of limestone. No tectonic structures are observed.

#### TRAVERSES SOUTH AND SOUTHEAST OF THE MOUNT ALBERT:

The traverses started from south of La Gite were continued south of the southern gate of the Gaspé National Park, northwest of the Ste. Anne Lake.

South of the La Gite sporadic roadside outcrops of the Shickshock metavolcanics are observed. They contain the  $S_1$  parallel to  $S_g$  layering and dip southwesterly, striking NW-SE.

South of the Isabelle Stream the Silurian Quartzite (Val Brilliant) is exposed, striking E-W, and dipping at  $55^\circ$ S.

Further south in outcrops of shaly, silty limestone of Cap Bon Ami Formation, the  $S_g$  trends N70E and dips steeply. As the traverse is continued further south the strike of the  $S_g$  swings to N-S. South of the Hog's Back Mountain the silty limestone is typically fractured by a consistent but coarse set of fractures trending at N70E. This set of fractures is supposed to be the  $S_2$  cleavage, owing to its regularity in development, and its trend which is typical of the  $S_2$  fracture in the region.

Half a mile west of the southern gate, shaly limestone is exposed. It dips at  $30^\circ$  and strikes at  $115^\circ$ . It is cut across by a steep fracture trending at N70E. The latter is considered as the  $S_2$  cleavage.

The rocks are poorly exposed. The Devonian limestone and sandstone are frequently mineralised and also intruded by dykes.

#### CONCLUSIONS.

1. Northeast and east of the Mount Albert, the trend of the  $F_1$  structures diverges from the  $F_2$  structures. The  $F_1$  trends NW-SE and the  $F_2$  continue in a ENE direction.

2. The  $S_1$  schistosity present in the grey phyllitic slates, and the conglomerates is refolded. This is absent from the Ordovician rocks exposed further north. The line that separates the Cambrian-

Ordovician rocks containing the  $F_1$  and  $F_2$  structures from the Ordovician, containing only the  $F_2$  structures probably lies between the localities 2/18 and 2/20. The observations on 2/19 are non-conclusive.

3. The northern contact between the metavolcanics and the interbedded grey slates is conformable. The interbedded slates have a baked appearance.

4. The  $F_4$  cross folds are irregular in their frequency of development even in the proximity of the ultrabasic intrusive. Thus they are not related to the igneous intrusion. They are therefore related to an independent cross folding.

5. The traverses south of the Mount Albert reveal that although the attitude of bedding swings to a north-south one, the cleavage still develops irregularly, but steeply and trends ENE. It is not rotated in the N-S position as the bedding is. It is therefore concluded that the N-S curvature in the Devonian sediments are due to N-S anisotropy and the structural trends in the Cambrian-Ordovician basement. The Devonian sediments were not compressed in the SE-NW direction due to the cross anisotropy of the basement, and thus were zones of relative tension. This probably has resulted in intrusion of small igneous bodies and also mineralisation in the area.

APPENDIX IV

## CONODONT ANALYSIS:

18 samples of shaly limestone and calcareous shale were collected from the N-S traverses north of the Shickshock Mountains, to test if conodonts could be used to ascertain the age of the belt of rocks of unknown age north of the Shickshock Mountains. No fossils were observed in the slates that contain the calcareous bands sampled, and none of the samples yielded any recognizable conodonts. Nonetheless it is believed that detailed conodont work may give positive results in younger rocks, the limestone interbeds in the Ordovician slates.

APPENDIX V.

## GRAPTOLITE FAUNA OF THE SLATE WEDGE IN THE SHICKSHOCK FAULT ZONE.

200 yards south of the confluence of the Wilson Brook and the Cap Chat River, a wedge of Ordovician slate is exposed along the Hammer Mill road. The dark grey to black micaceous slate is tightly folded and a fracture cleavage develops parallel to the axial plane of the fold. The fold is regarded as  $F_2$  and cleavage the  $S_2$ . The rock is low in metamorphic grade and dull in lustre.

Graptolites usually deformed and flattened, observed on the  $S_s$  surface, were identified by Dr. B. Erdtman of Carleton University, and are listed below:

Climacograptus wilsoni cf. tubularis ELLES & WOOD

Climacograptus cf. C. brevis ELLES & WOOD

Orthograptus cf. O. truncatus LAPWORTH

Orthograptus cf. O. micracanthus ELLES & WOOD

Glyptograptus sp. or cf. Climacograptus brevis ELLES & WOOD

The above fossils sufficiently bracket the age of the slate as Middle Caradocian, Middle Ordovician.

Mattinson (1964, p. 69) examined the graptolites from the locality. He suggested early Middle Ordovician age for the slates.

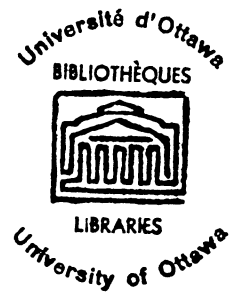
SC

VOLUME II

STRUCTURAL ANALYSIS  
OF THE LOWER PALEOZOIC  
ROCKS OF WESTERN GASPE, QUEBEC

ILLUSTRATIONS

( PLATES AND PHOTOGRAPHS )



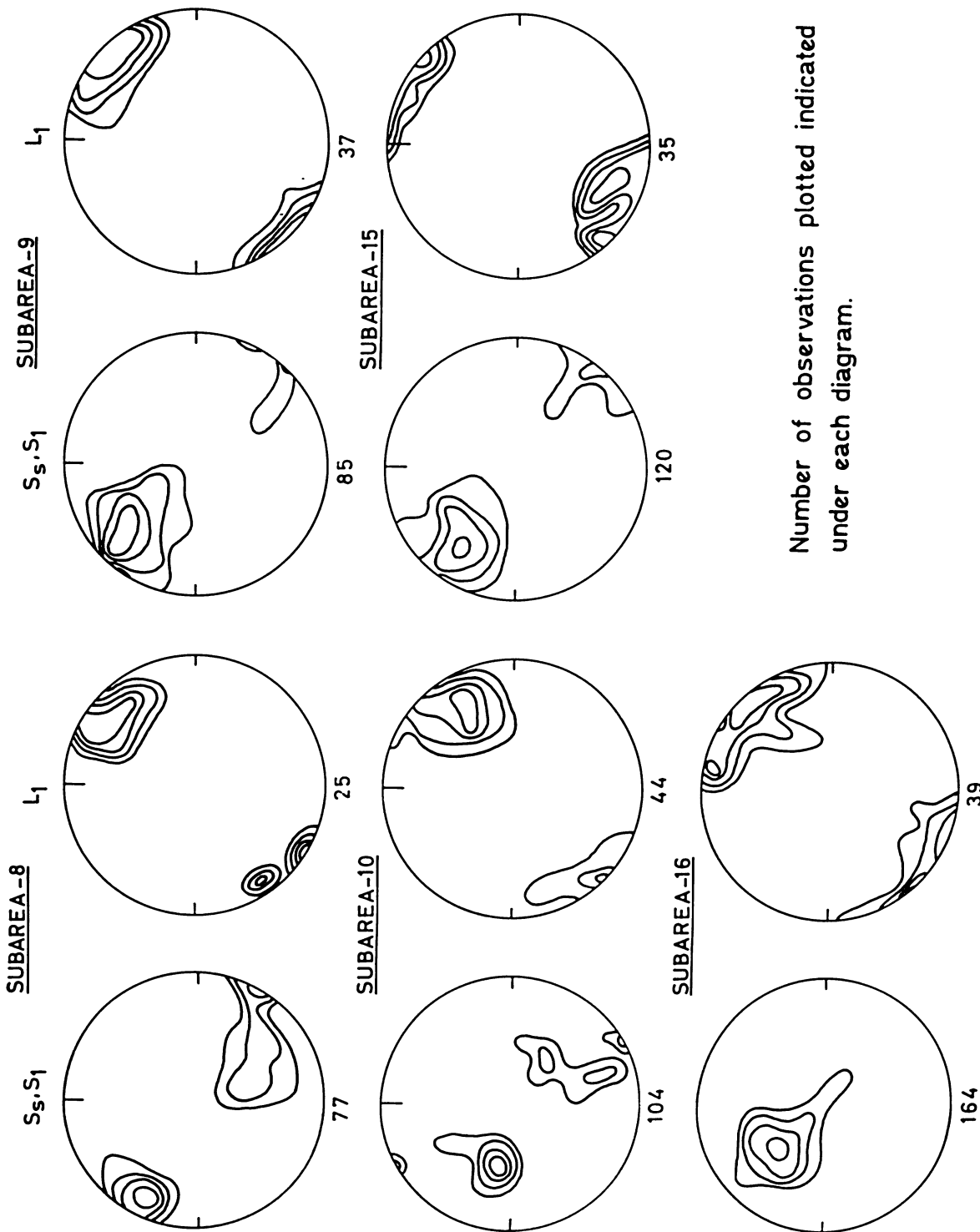
PLATES

PLATE I

Net projections of  $F_1$  structures in Cambrian-Ordovician rocks. The polar plots on equal area nets are contoured at 2, 4, 6, 10 per cent frequencies per 1% area. The figures given under projections are number of observations.

PLATE 1

**F<sub>1</sub> STRUCTURES IN CAMBRIAN-ORDOVICIAN ROCKS**  
**Contoured on Equal-Area Polar Net for Each Subarea (fig.9)**



Number of observations plotted indicated  
under each diagram.

## PLATE II

F<sub>2</sub> fold profiles showing variation in bed thickness: Most examples are sampled or photographed looking down the plunge in Cambrian and Ordovician slate interbedded with calcareous siltstone or limestone from the St. Lawrence coast. In d, 1, 2 and 3 folds are from Shickshock metavolcanics.

Under each drawing of the fold, the curves represent the thicknesses plotted against the reference lines: 't' is the thickness normal to bedding, and 'T' is the thickness parallel to axial plane. The thicknesses between reference lines are approximate.

The reference lines are drawn at limb and hinge areas of the folds. Usually they are approximate isogon lines, but for sake of clarity, occasionally they are not. Although the distance between two reference lines is usually variable, to avoid tedious construction, all reference lines on a sample are drawn at equal distances for thickness relationships.

PLATE IIa

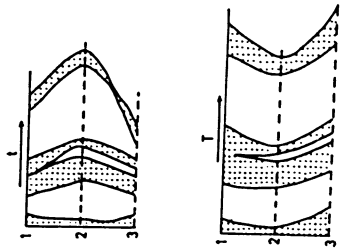
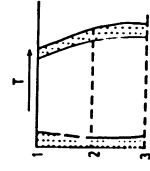
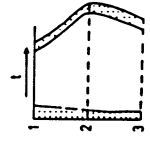
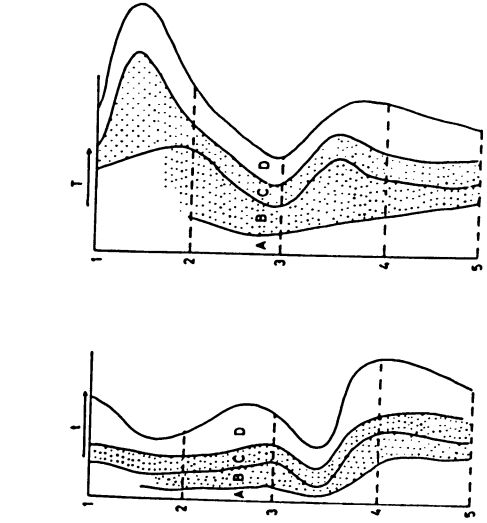
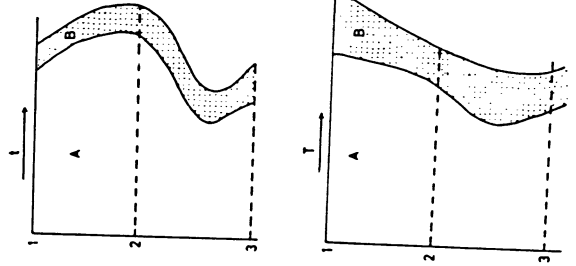
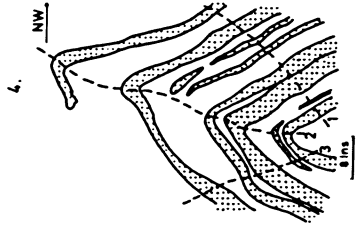
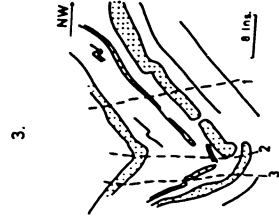
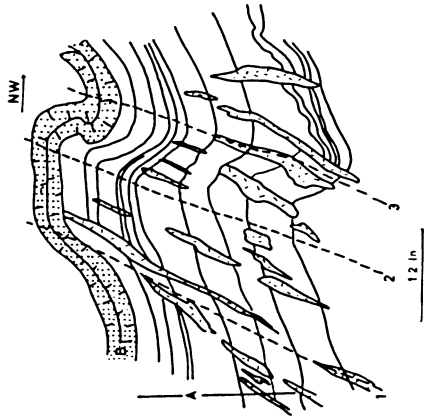
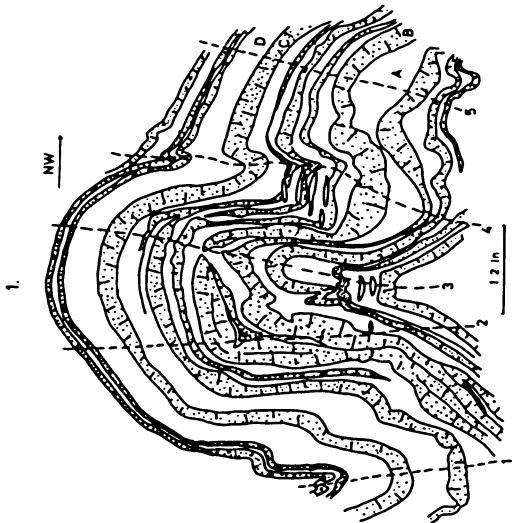


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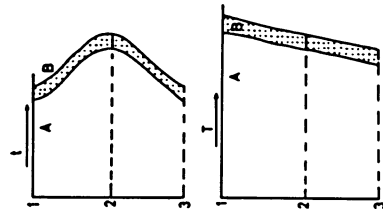
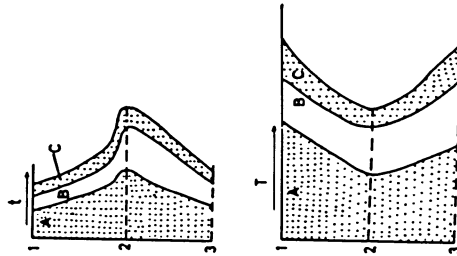
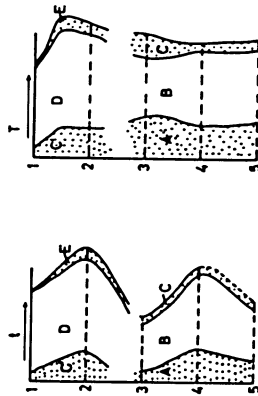
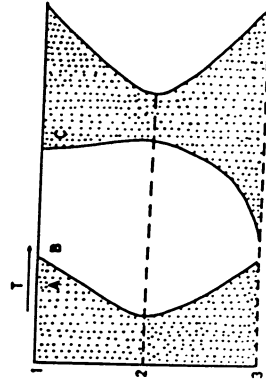
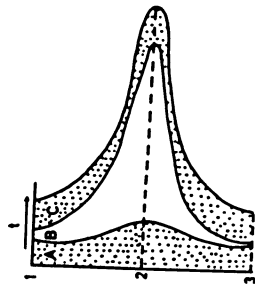
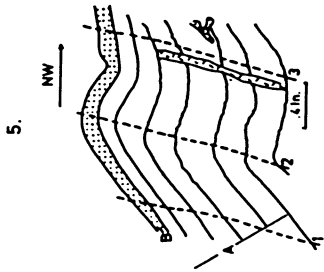
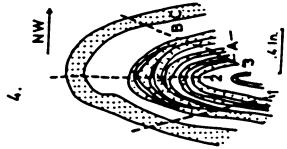
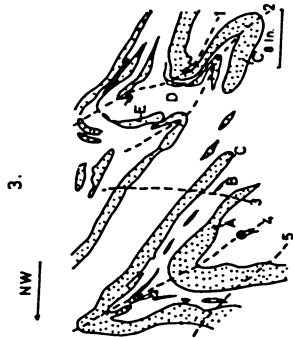
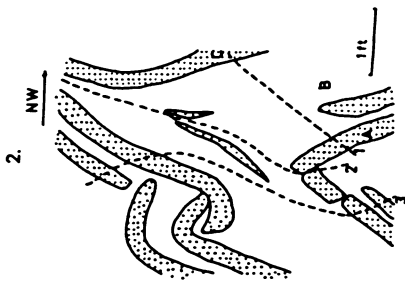
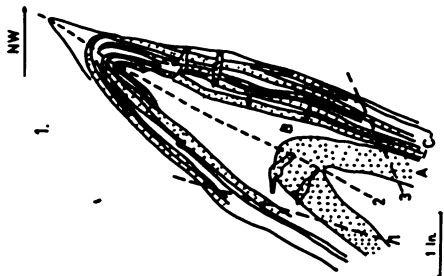


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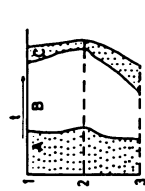
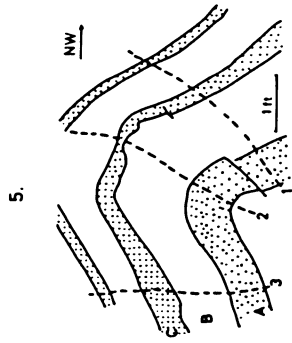
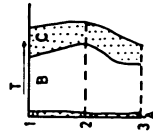
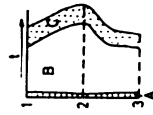
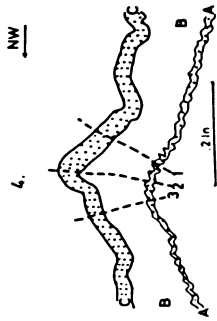
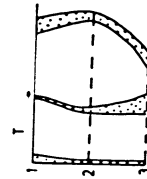
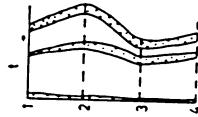
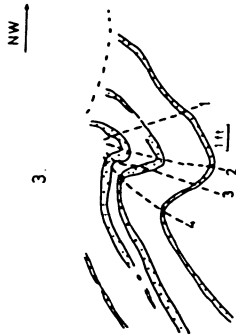
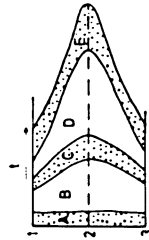
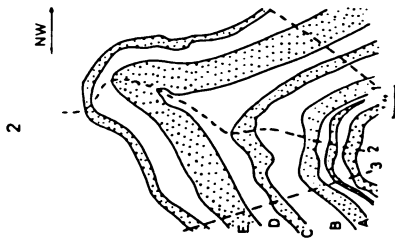
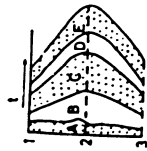
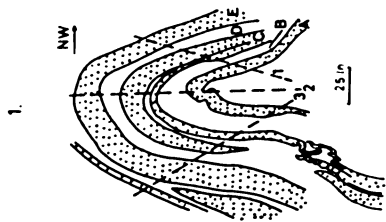


PLATE III d

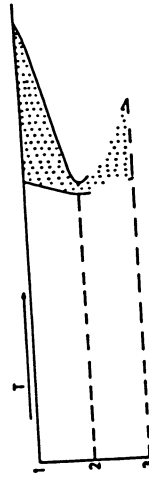
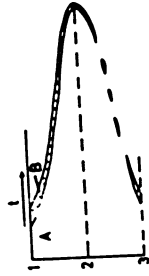
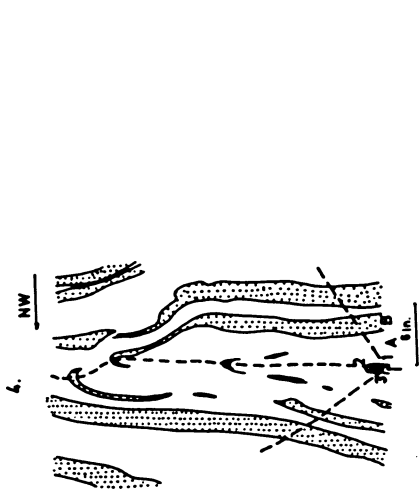
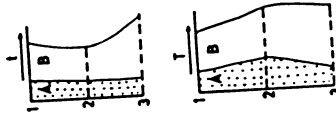
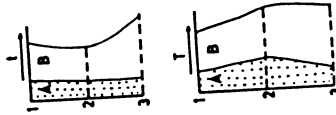
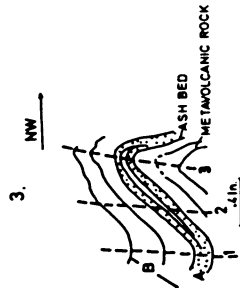
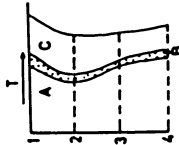
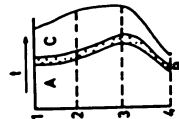
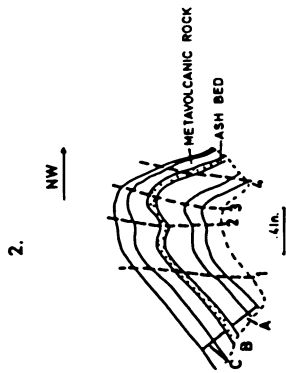
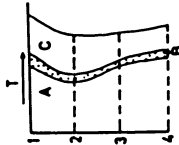
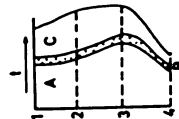
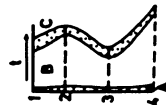
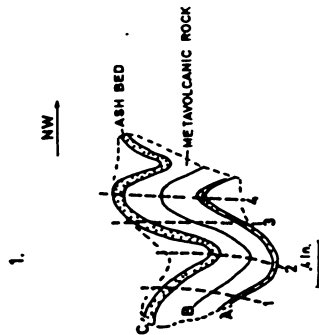


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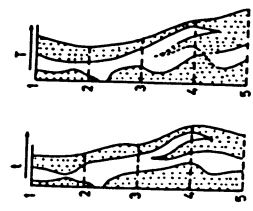
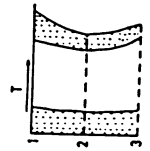
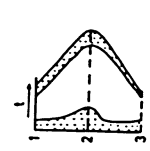
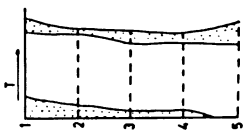
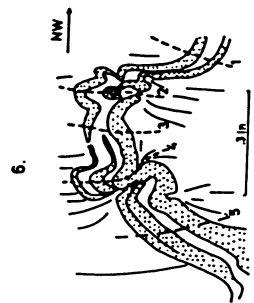
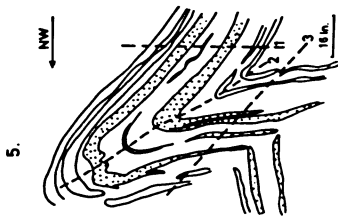
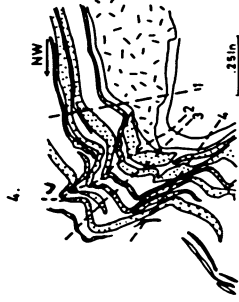
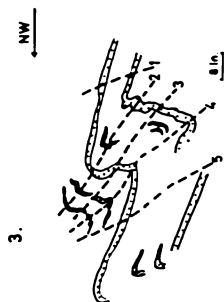


PLATE III

Net projections of  $F_2$  structures in Lower Paleozoic rocks. The polar plots on equal area nets are contoured at 2, 4, 6 and 10 percent frequencies per 1% area. The figures given under diagrams are the number of observations.

PLATE IIIa

**F<sub>2</sub> STRUCTURES IN LOWER PALEOZOIC ROCKS**  
Contoured on Equal-Area Polar Net for Each Subarea (fig.2)

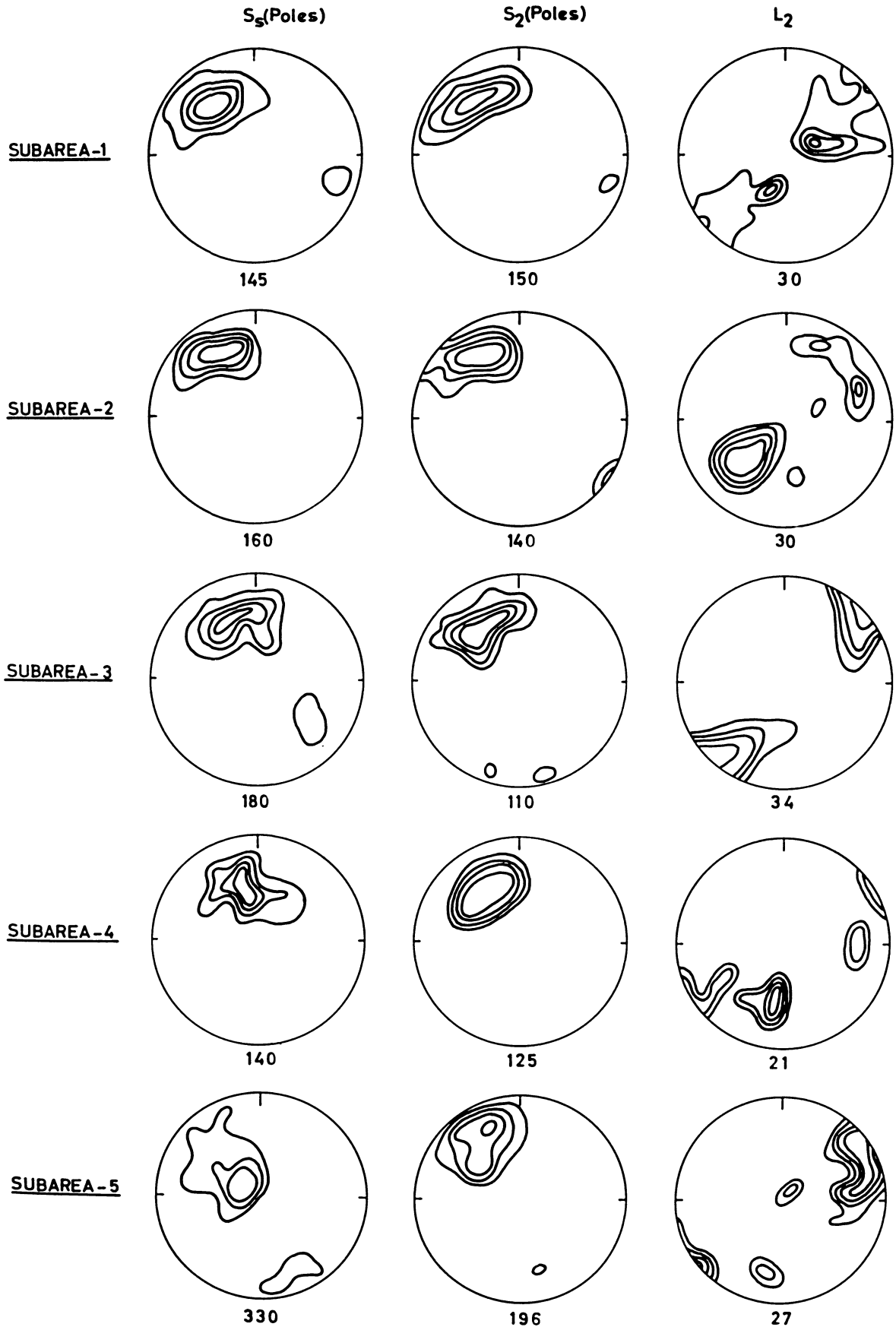


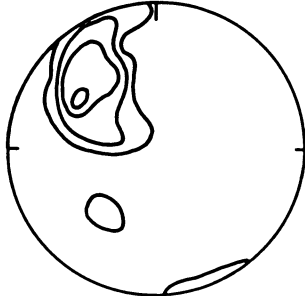
PLATE IIIb

$S_5$ (Poles)

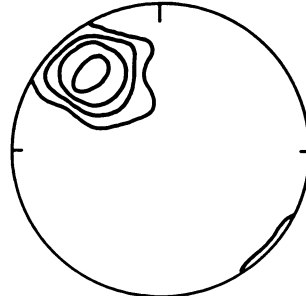
$S_2$ (Poles)

$L_2$

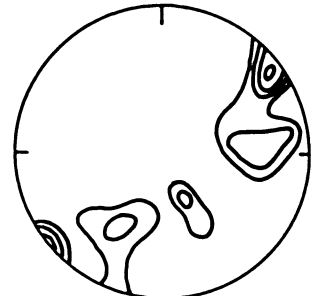
SUBAREA-6



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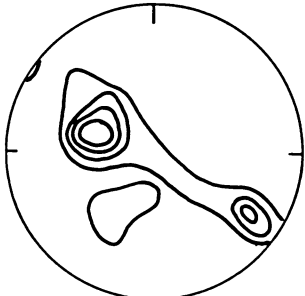


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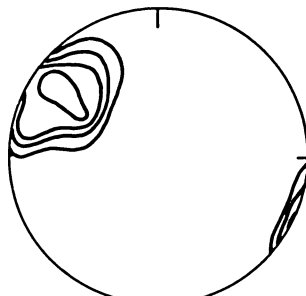


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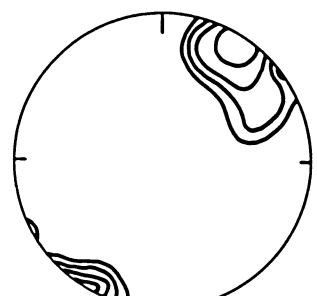
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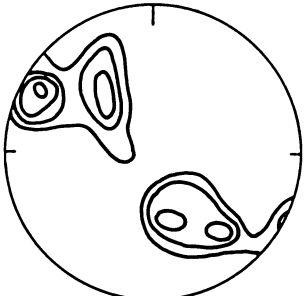


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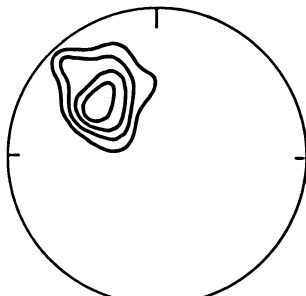


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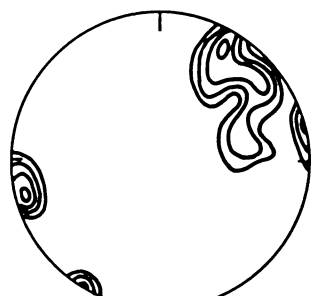
SUBAREA-8



126

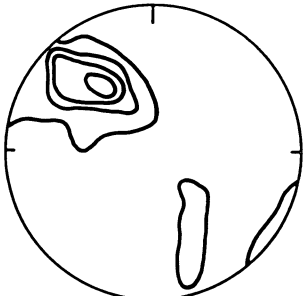


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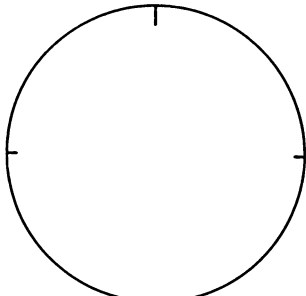


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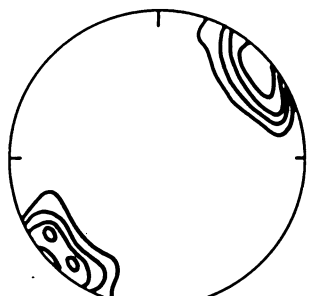
SUBAREA-9



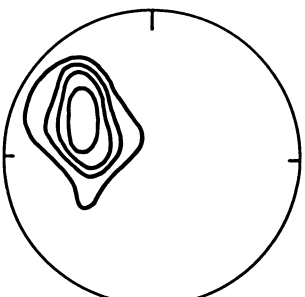
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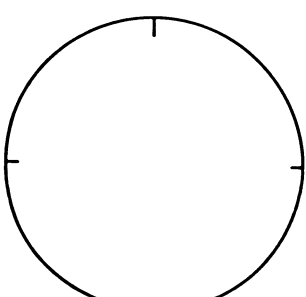
85



SUBAREA-10



83



25

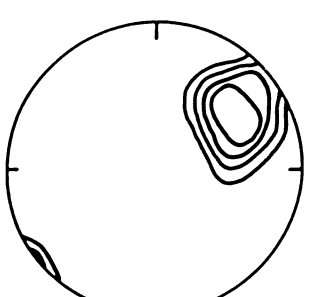


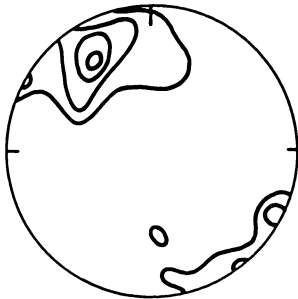
PLATE IIIc

$S_5(\text{Poles})$

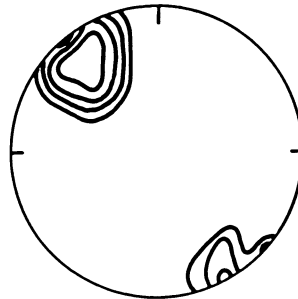
$S_2(\text{Poles})$

$L_2$

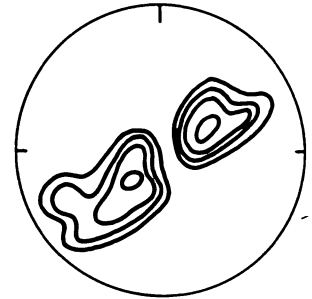
SUBAREA-11



143

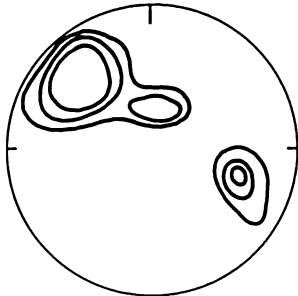


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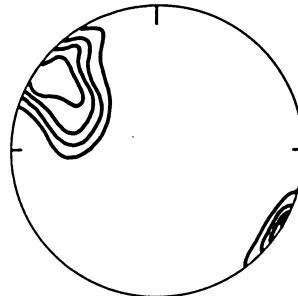


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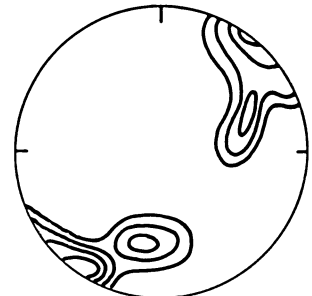
SUBAREA-12



175

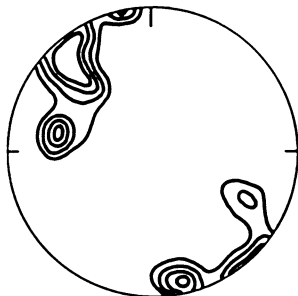


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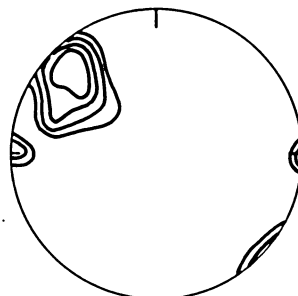


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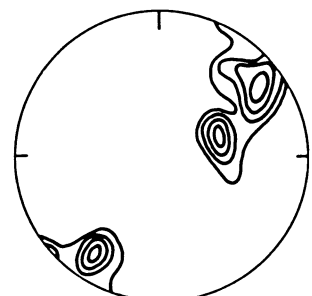
SUBAREA-13



42

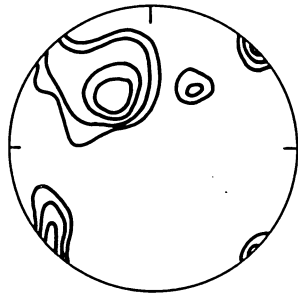


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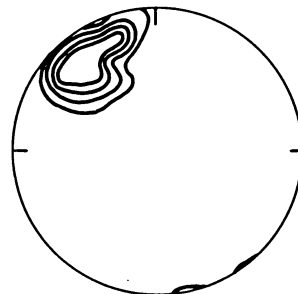


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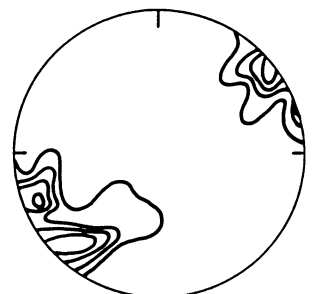
SUBAREA-14



38

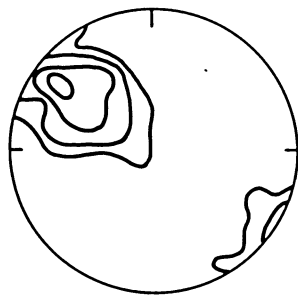


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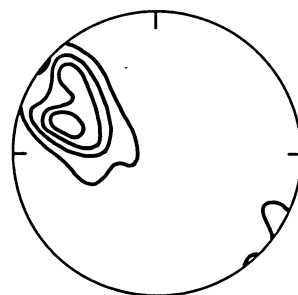


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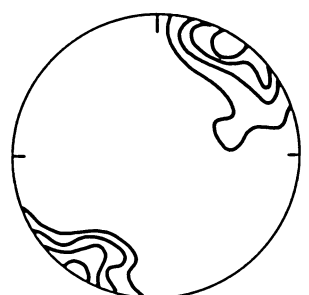
SUBAREA-15



138



165



110

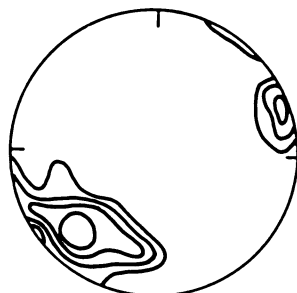
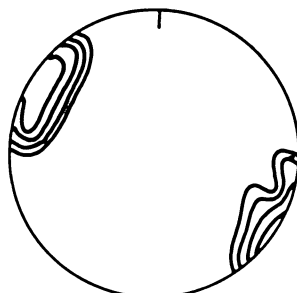
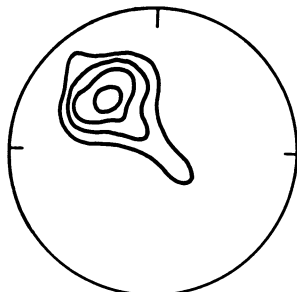
PLATE III d

$S_5$  (Poles)

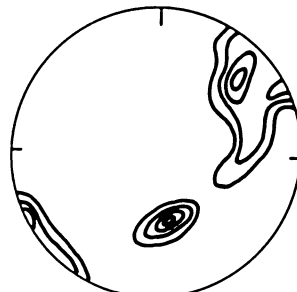
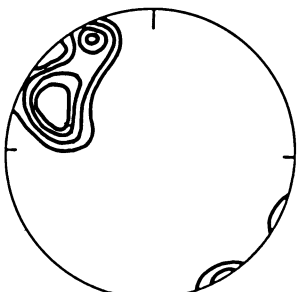
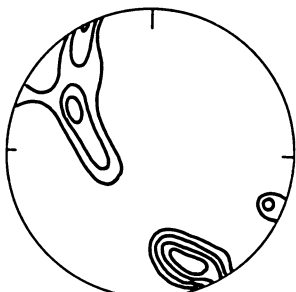
$S_2$  (Poles)

$L_2$

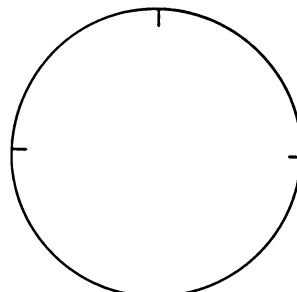
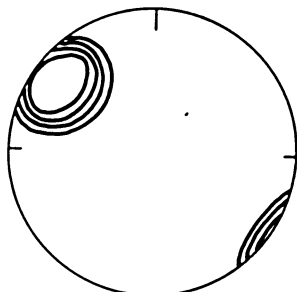
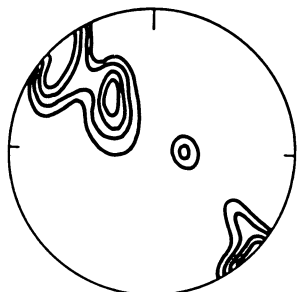
SUBAREA -16



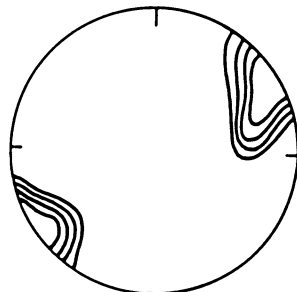
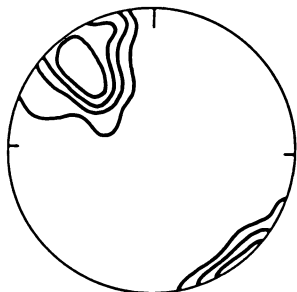
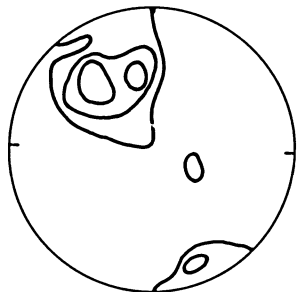
SUBAREA -18



SUBAREA -19, 20



SUBAREA -21



SUBAREA -22

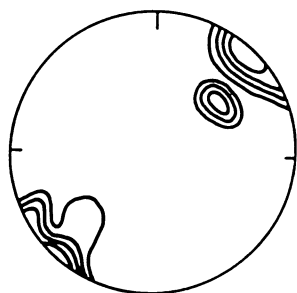
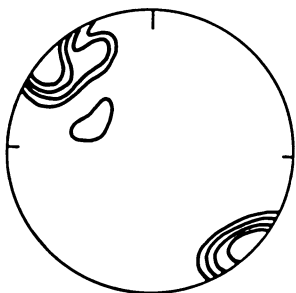
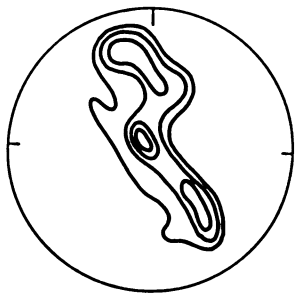


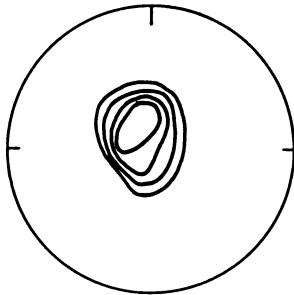
PLATE - IIIe

$S_5$ (Poles)

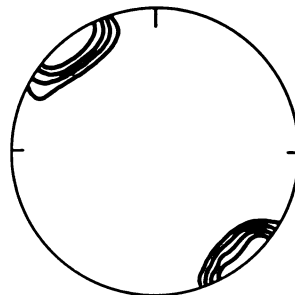
$S_2$ (Poles)

$L_2$

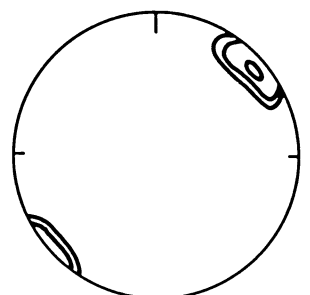
SUBAREA-23



24

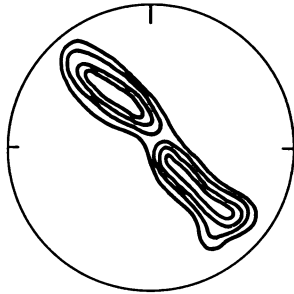


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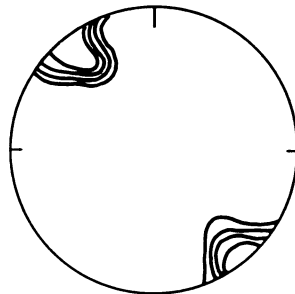


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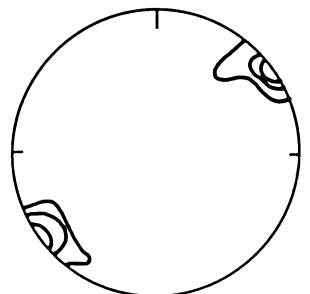
SUBAREA-24



20

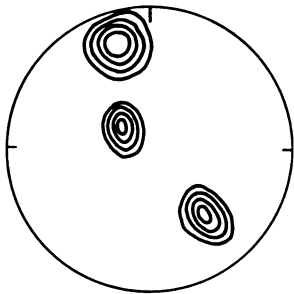


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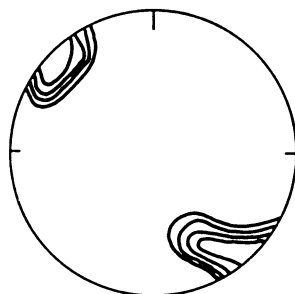


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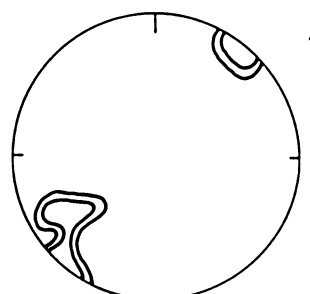
SUBAREA-25



22

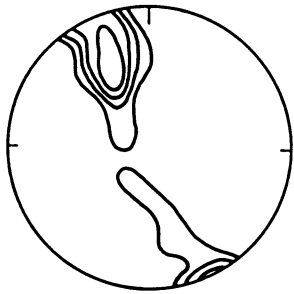


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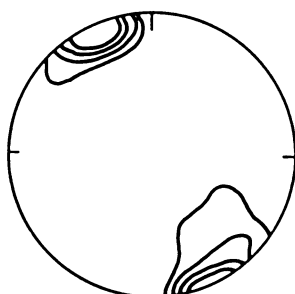


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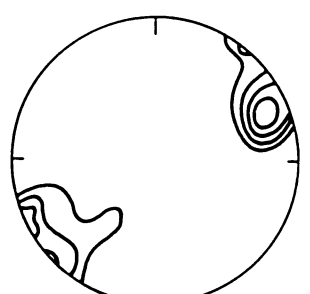
SUBAREA-26



90

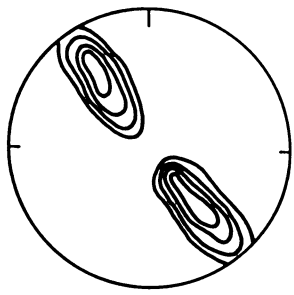


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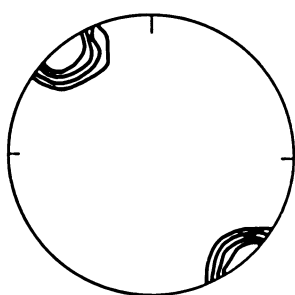


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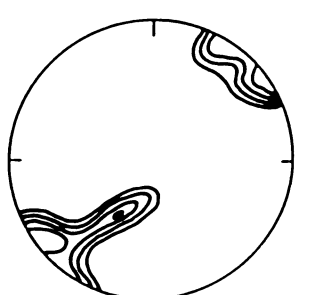
SUBAREA-27



40



80



16

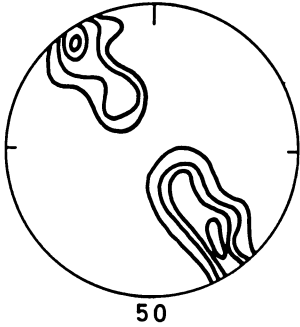
PLATE III f

$S_5$ (Poles)

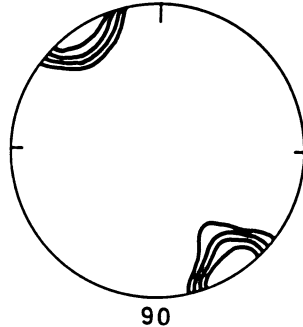
$S_2$ (Poles)

$L_2$

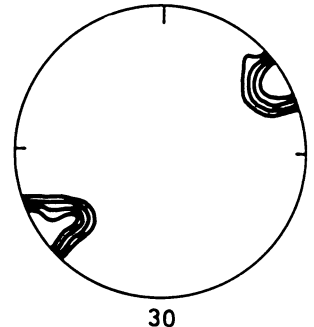
SUBAREA-28



50

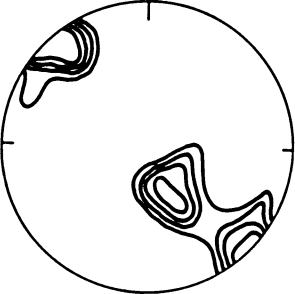


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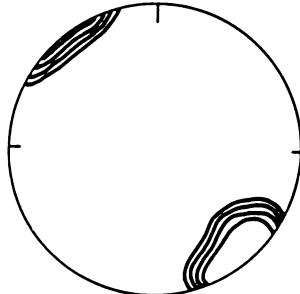


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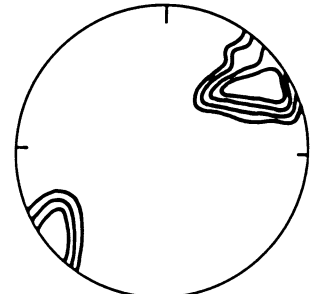
SUBAREA-29,30



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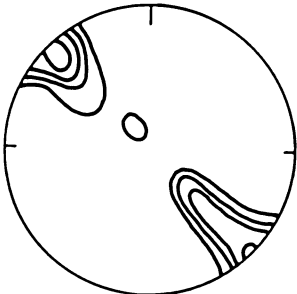


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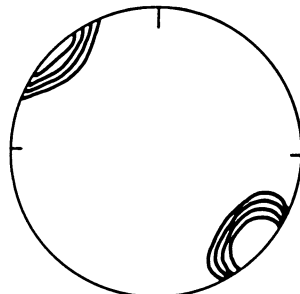


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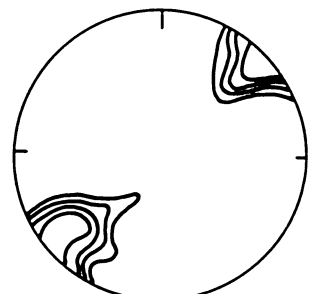
SUBAREA-31



87

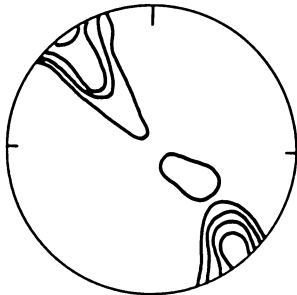


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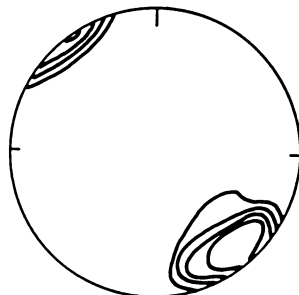


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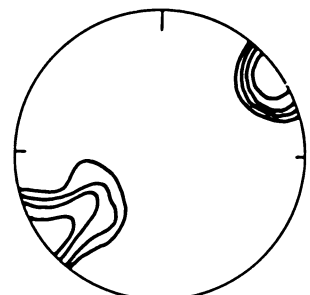
SUBAREA-32



115

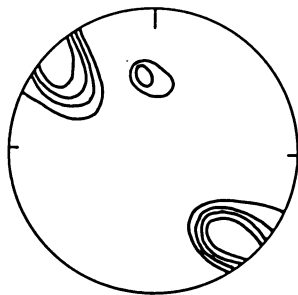


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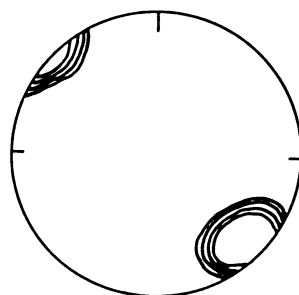


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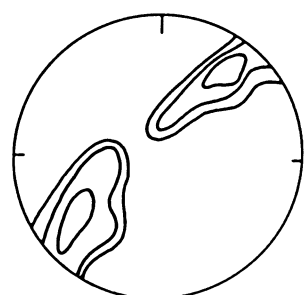
SUBAREA-33



110

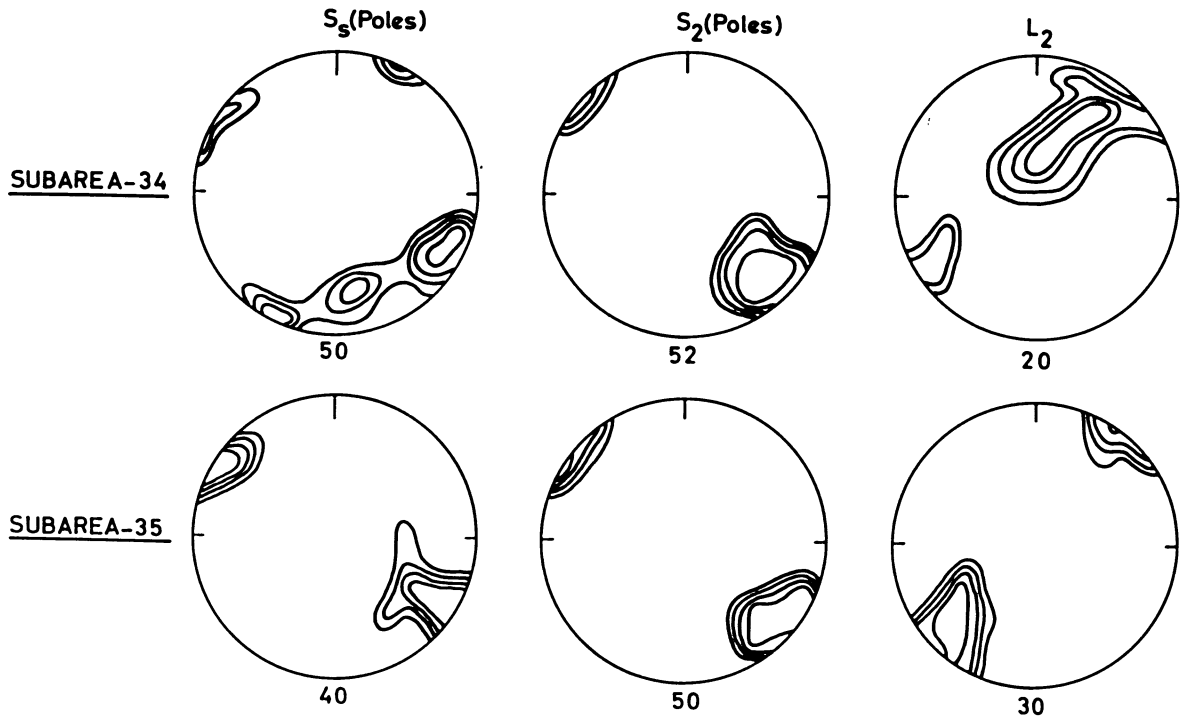


90



66

PLATE IIIg



Number of observations plotted indicated under each diagram.

PLATE IV

Net projections of  $F_2$  structures in Lower Paleozoic rocks to demonstrate geometrical relationships:

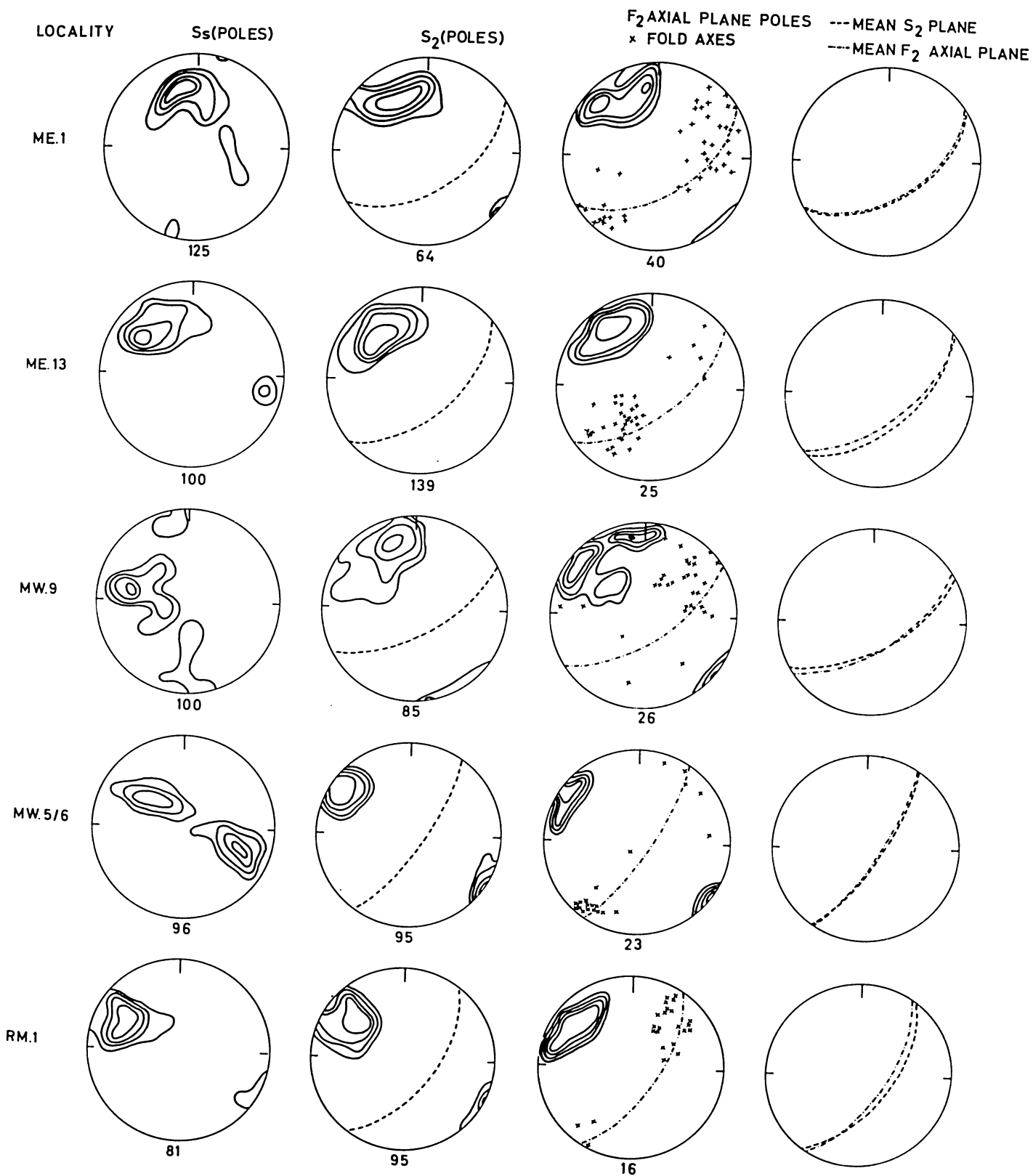
- a. Plotted for localities in Cambrian-Ordovician and Ordovician rocks to demonstrate coincidence of mean  $S_2$  cleavage and  $F_2$  axial plane.
- b. Plotted for subareas to demonstrate relationship between  $L_2$  and  $F_2$  fold axes in Cambrian-Ordovician and Ordovician rocks. Dashed great circles define modal  $F_2$  axial planes.
- c. Plotted for Upper-Ordovician and Siluro-Devonian rock formations or groups. Dashed great circles define modal  $F_2$  axial planes.

The polar plots are contoured at 2, 4, 6, and 10 percent frequencies per 1% area. The figures given under diagrams are the numbers of observations.

PLATE IVa

F<sub>2</sub> STRUCTURES IN CAMBRIAN AND ORDOVICIAN ROCKS

Contoured on Equal-Area Polar Net



LOCALITIES SHOWN IN FIGURE 2.

PLATE IVb

**F<sub>2</sub> STRUCTURES IN CAMBRIAN AND ORDOVICIAN ROCKS**  
Contoured on Equal-Area Polar Net. Coincidence of L<sub>2</sub> and  
Observed F<sub>2</sub> Fold Axes in Subareas is Demonstrated.

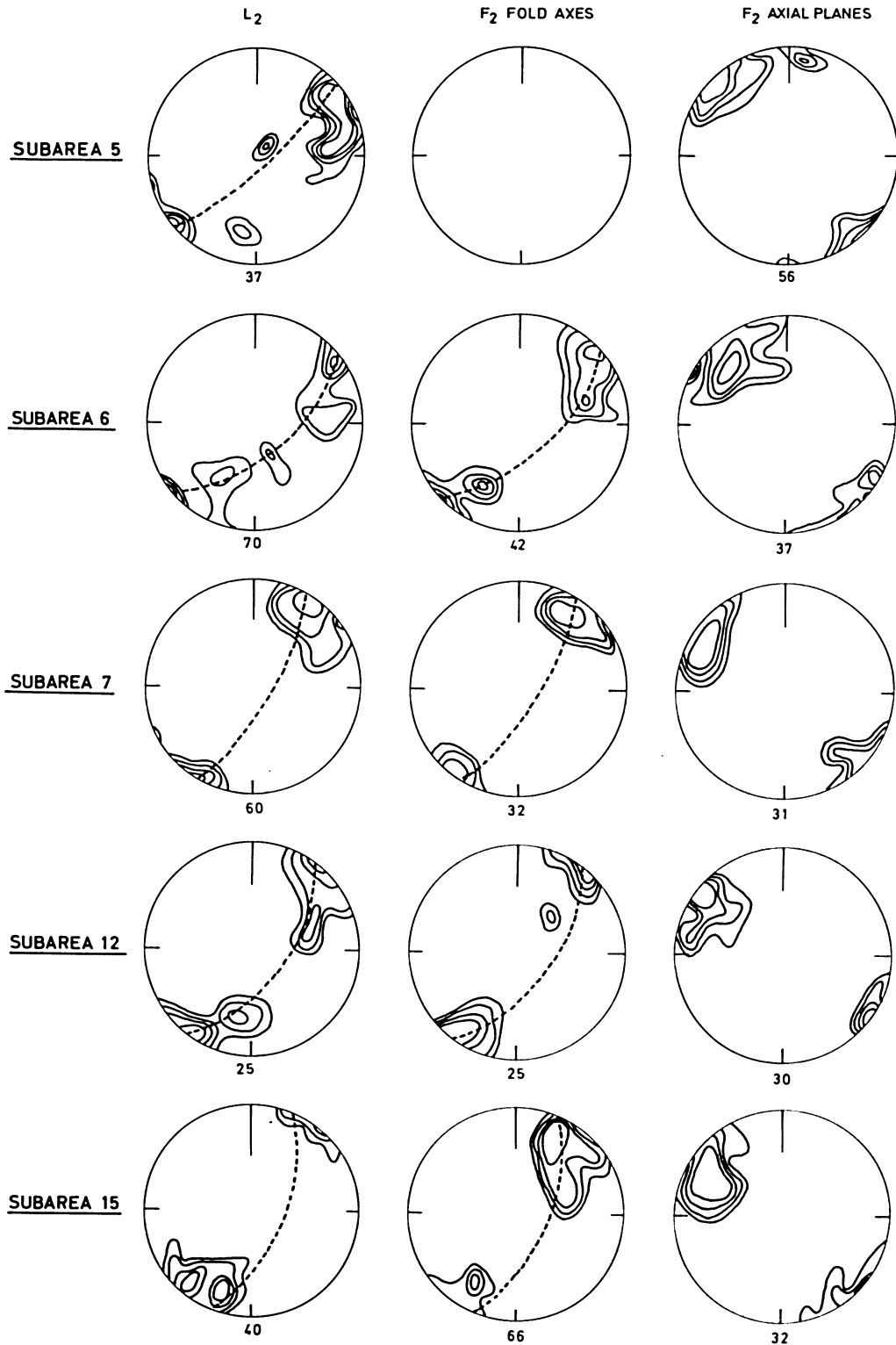


PLATE IVc

F<sub>2</sub> STRUCTURES IN UPPER ORDOVICIAN AND SILURO-DEVONIAN ROCKS  
Contoured on Equal-Area Polar Net

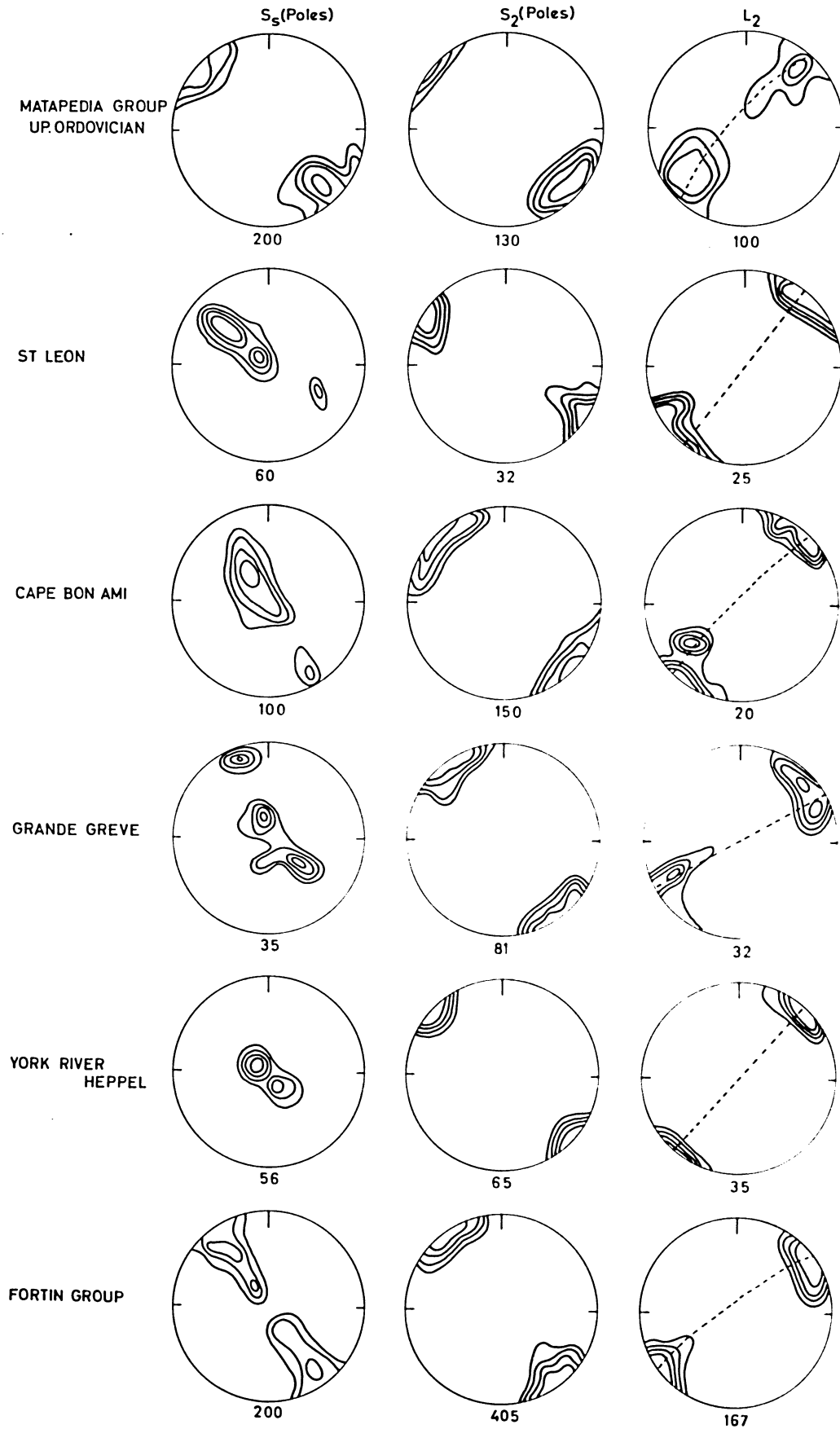
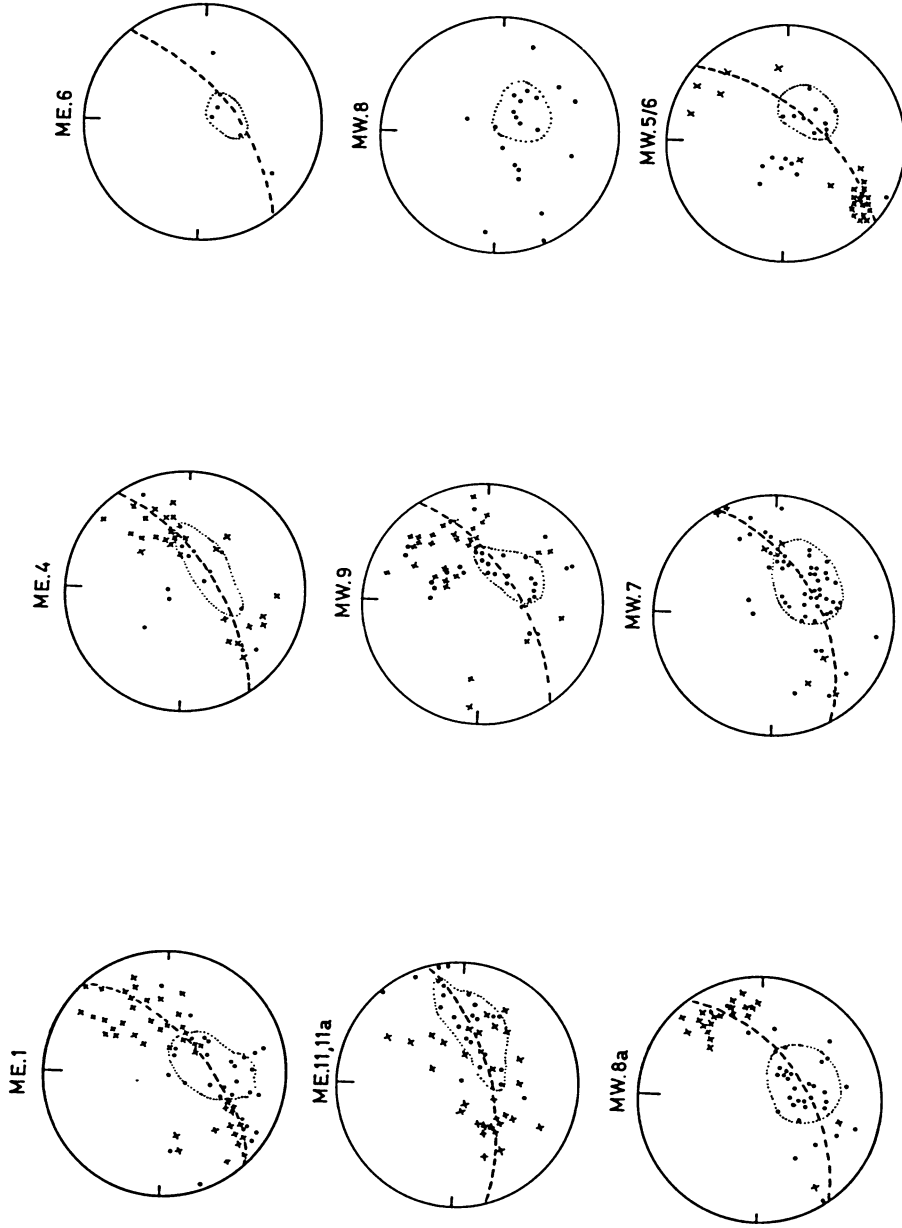


PLATE V

Net projections of slickensides on calcite veins plotted on Wulff net for each locality. The Wulff net was used for improved clarity of the low plunging plots. Dotted line encloses the predominant slickenside set.

PLATE V

SLICKENSIDE PLUNGES ON CALCITE VEINS  
Plotted on Wulff Net with  $F_2$  Axial Plunges for Each Locality



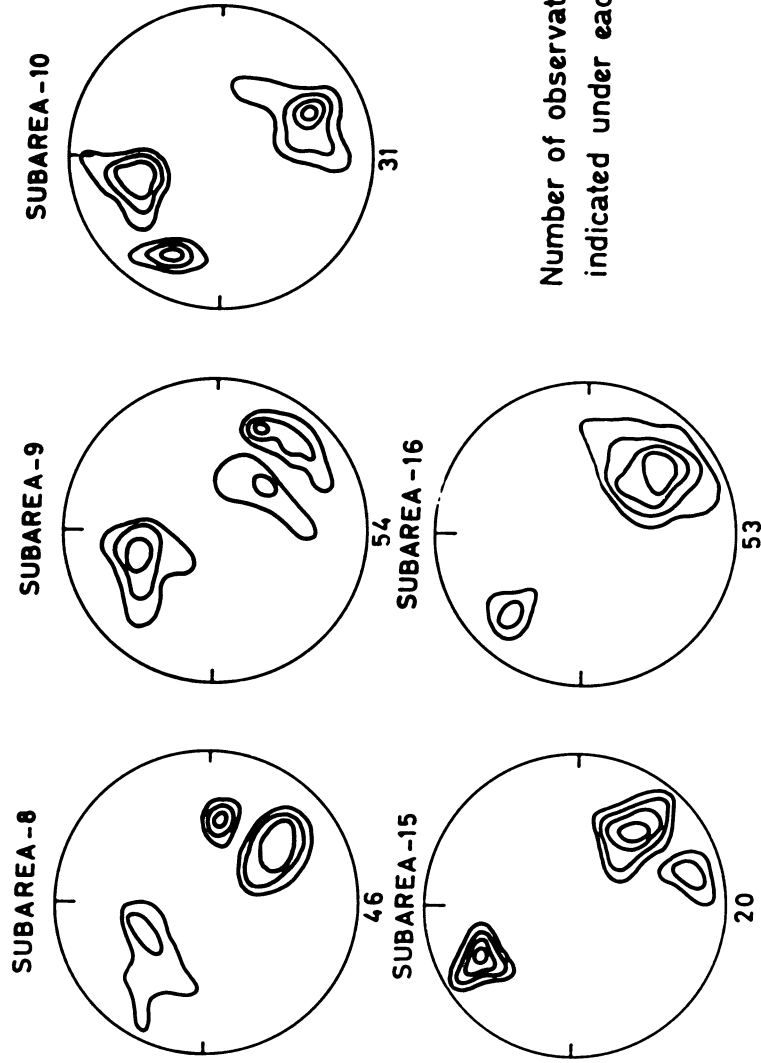
- SLICKENSIDE PLUNGE
  - x  $F_2$  AXIAL PLUNGE
- DASHED GREAT CIRCLE INDICATES APPROXIMATE MEAN AXIAL PLANE FOR EACH LOCALITY  
CLUSTER OF SLICKENSIDE PLUNGES ENCLOSED IN FINE DOTTED LINE.  
HEADINGS (ME.1) ARE LOCALITIES SHOWN IN FIG.2.

PLATE VI

Net projections of  $F_4$  structures in Cambrian-Ordovician rocks for subareas (see fig. 9). The projections are contoured for 2, 4, 6, and 10 percent frequencies per 1% area. The figures given under diagrams are the number of observations plotted.

PLATE VI

**F<sub>4</sub> STRUCTURES IN CAMBRIAN-ORDOVICIAN ROCKS**  
**L<sub>4</sub> Contoured on Equal-Area Polar Net for Each Subarea (fig.9)**



Number of observations plotted indicated under each diagram.

PLATE VII

Net projections of joint pattern in Lower Paleozoic rocks. Joint poles contoured for 2, 4, 6 and 10 percent frequencies per 1% area.

a. Joint pattern in Cambrian and Ordovician rocks plotted for each locality, the number of which is given under each diagram. Figure in brackets given under each diagram is the number of observations plotted.

b. Joint pattern in Siluro-Devonian rocks plotted for rock units. Figures given in brackets are the numbers of observations plotted.

JOINT PATTERN IN CAMBRIAN AND ORDOVICIAN ROCKS  
 Poles Contoured on Equal-Area Polar Net for Each Locality (Fig. 21)

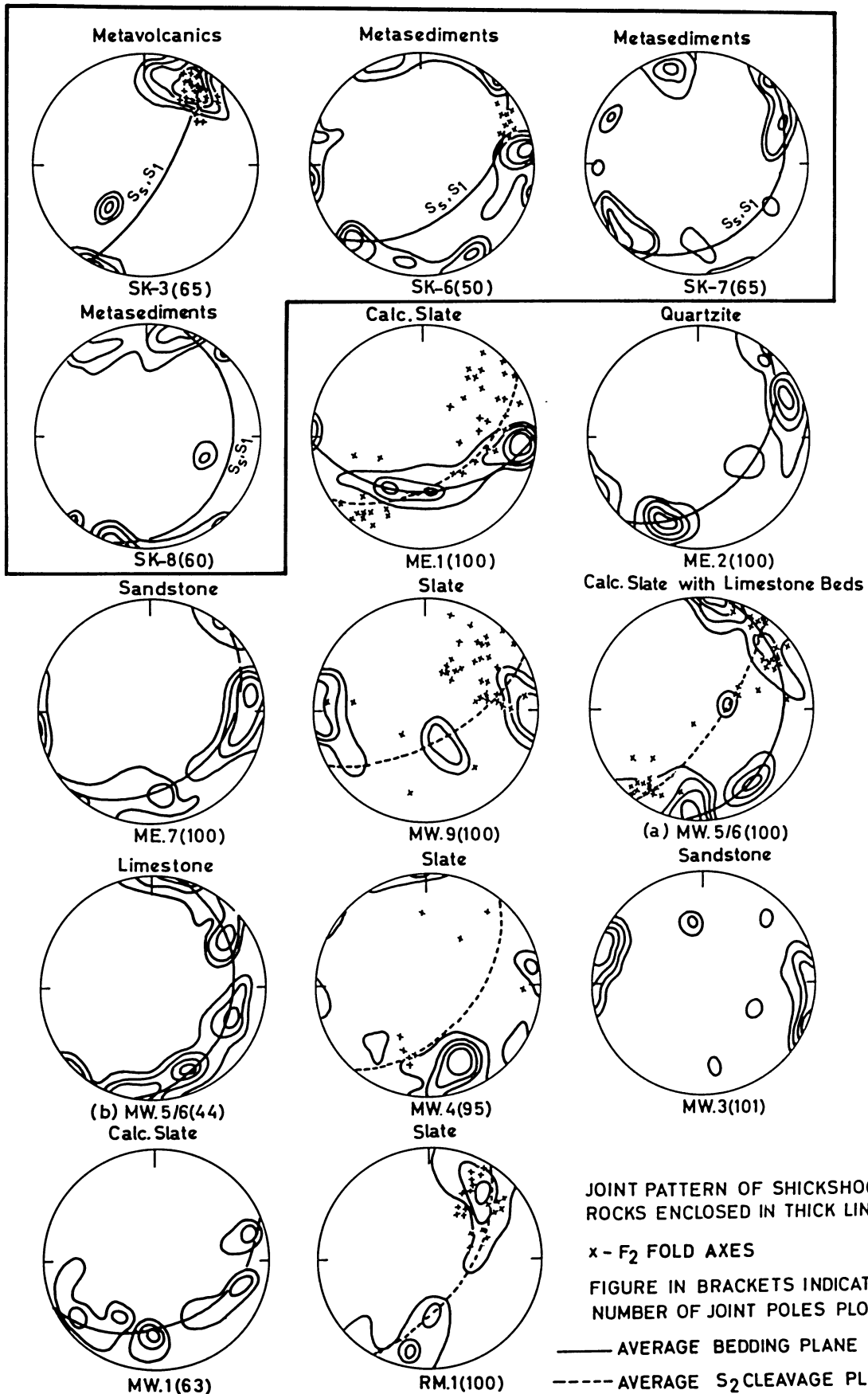
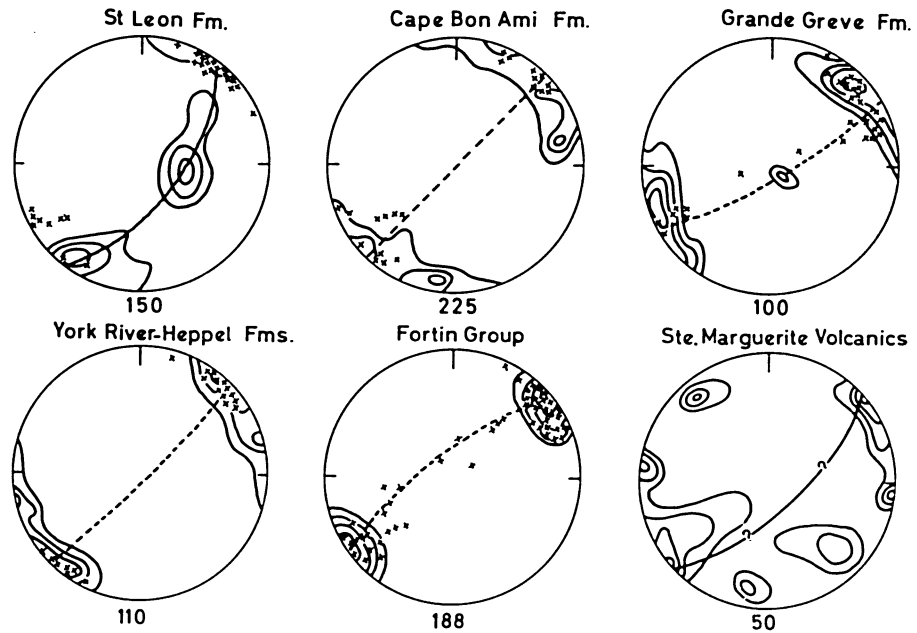


PLATE VIIb

JOINT PATTERN IN SILURO-DEVONIAN ROCKS

Poles Contoured on Equal-Area Polar Net.



x F<sub>2</sub> FOLD AXES

FIGURE GIVEN UNDER EACH DIAGRAM INDICATES  
THE NUMBER OF OBSERVATIONS

— AVERAGE BEDDING PLANE

- - - AVERAGE S<sub>2</sub> CLEAVAGE PLANE

GEOLOGICAL PHOTOGRAPHS

## Notes On the geological photographs

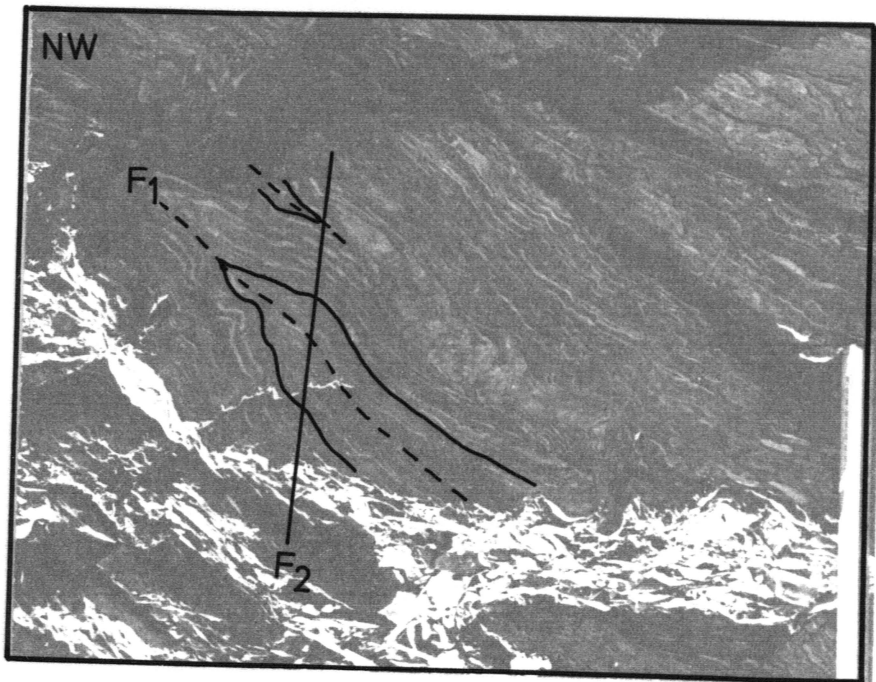
The photographs in field were taken with a 35mm camera, using Plus X black and white and Kodachrome II colour transparency film.

The field photographs, which show prominent grain in print, have been printed from black and white negatives made from colour transparencies. Best results were obtained by projecting the transparency slide on a white screen, and rephotographing the projection with a 135 mm telephoto lens.

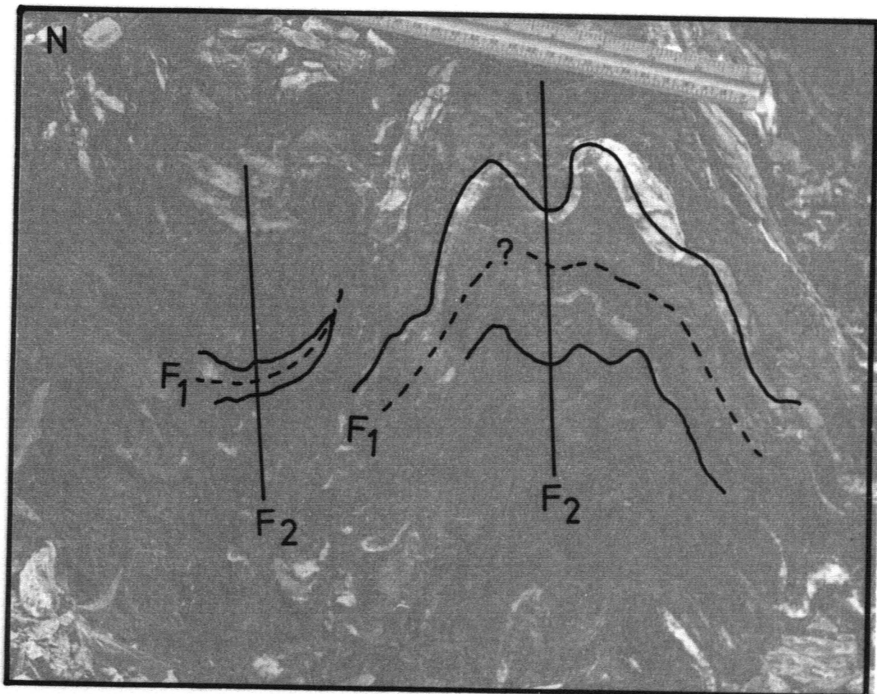
The photographs 57, and between 94 and 100 were obtained by making a orthochromatic negative by putting the rock thin section directly into 35 mm photographic enlarger.

Extensive use of red and yellow filter was made to bring out the contrast in colour of the rocks while photographing and rephotographing in black and white.

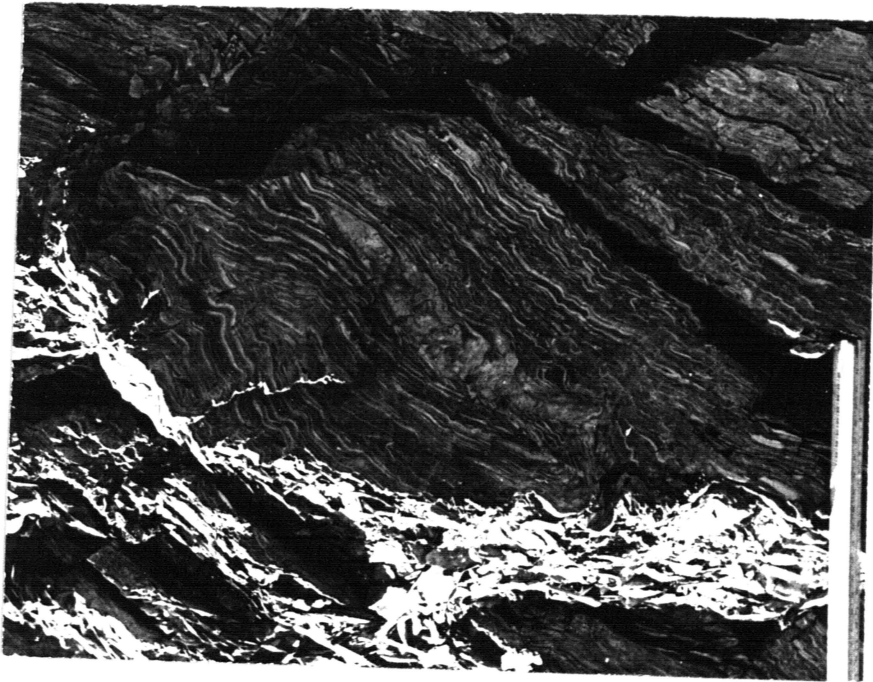
The graduation of scale in the photographs of the hand specimens is in inches.



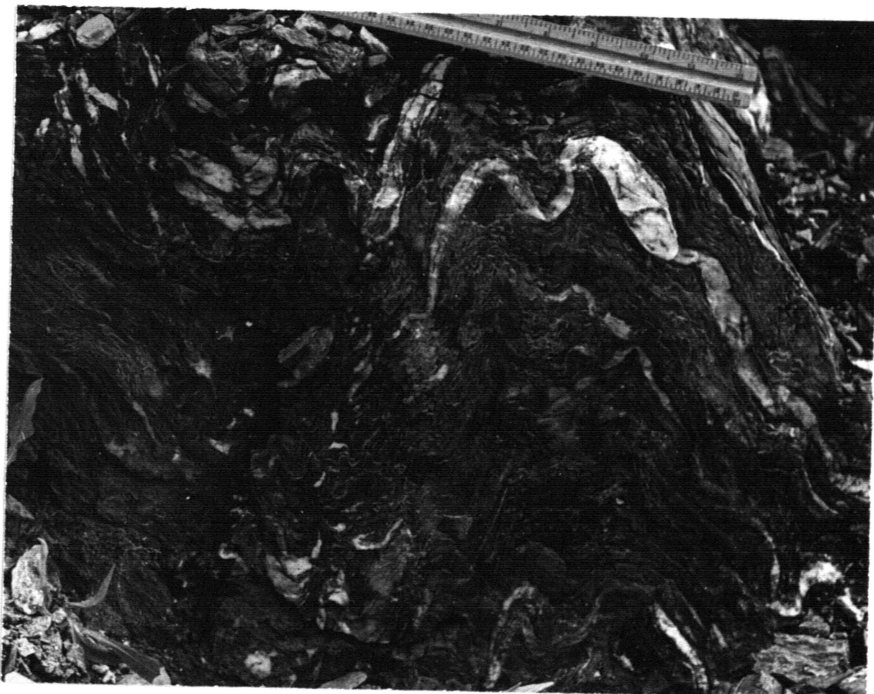
Photograph  
1.



Photograph  
2.



Photograph  
1.



Photograph  
2.

Photograph 3.

Tight inclined  $F_1$  (?) anticline and syncline. Boudins developed on the left synclinal limb (on the left of the field assistant) in grey calcareous slate. Locality 100 yards north of Riviere Matane village.

Photograph 4.

Chlorite lensoids on  $S_1$  schistosity surface in the Shickshock metavolcanics. Locality: East of Mount Albert.



Photograph  
3.



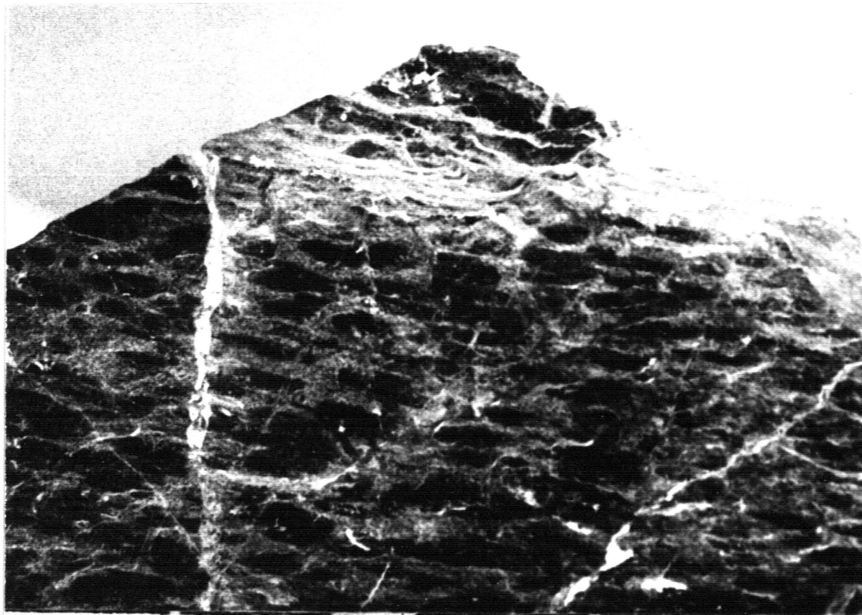
Photograph  
4.

Photograph 5.

Chlorite lensoids in the Shickshock metavolcanics.

Photograph 6.

$S_1$  schistosity surface of Cambrian-Ordovician grey slate with  $L_1(S_s/S_1)$  lineation, obliquely crossed by pitching  $L_2(S_1/S_2)$  and crenulation lineation. Locality: The Shickshock Mountains, 3 miles east of Depot Jean.



Photograph  
5.

4 5 6 7 8

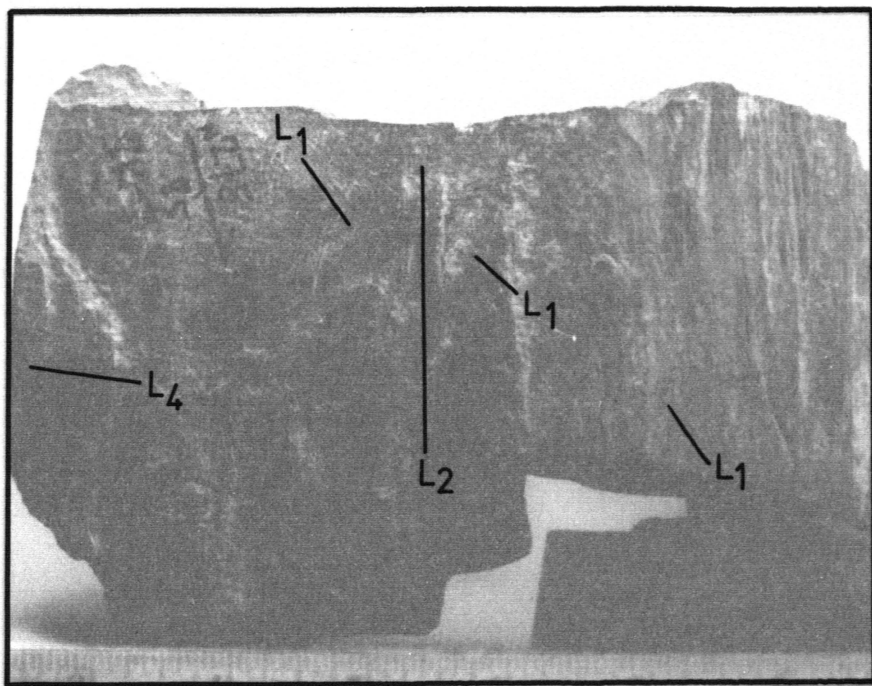


Photograph  
6.

4 5 6 7 8 9 10

Photograph 7.

F<sub>2</sub> fold deforming S<sub>1</sub> schistosity, L<sub>2</sub> crenulation lineation along the hinge area. A faint L<sub>1</sub> also is present oblique to the L<sub>2</sub>. Locality: SK 3, Shickshock Mountains.



Photograph  
7.



Photograph  
7.

Photographs 8 and 8a.

Subhorizontal foliation probably  $S_1$  in inclined  
Cambrian-Ordovician grey slates. Locality:  
(Appendix III) St.A 2/13, NE of Mount Albert,  
along the Trans Gaspé Highway.



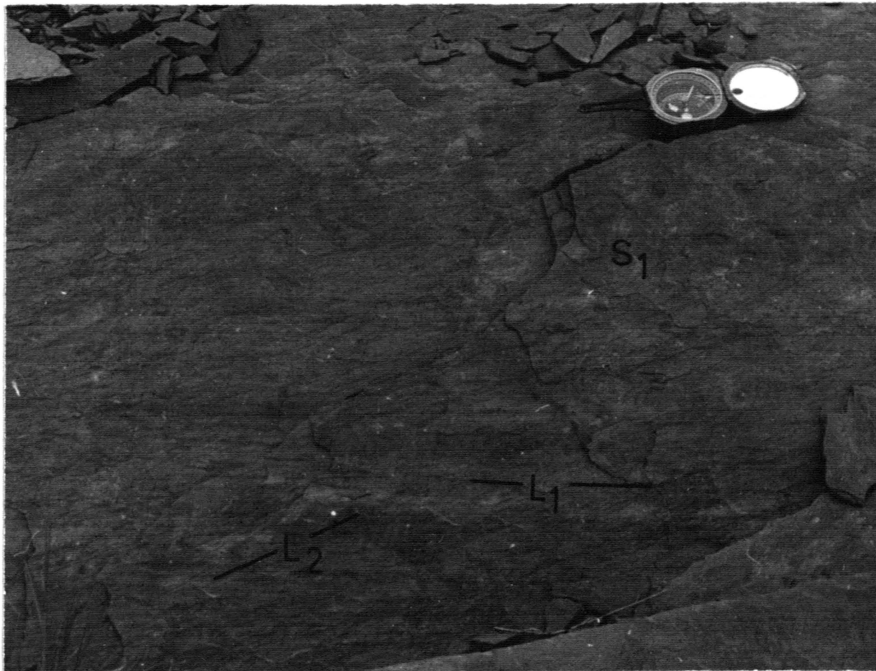
Photograph  
8.



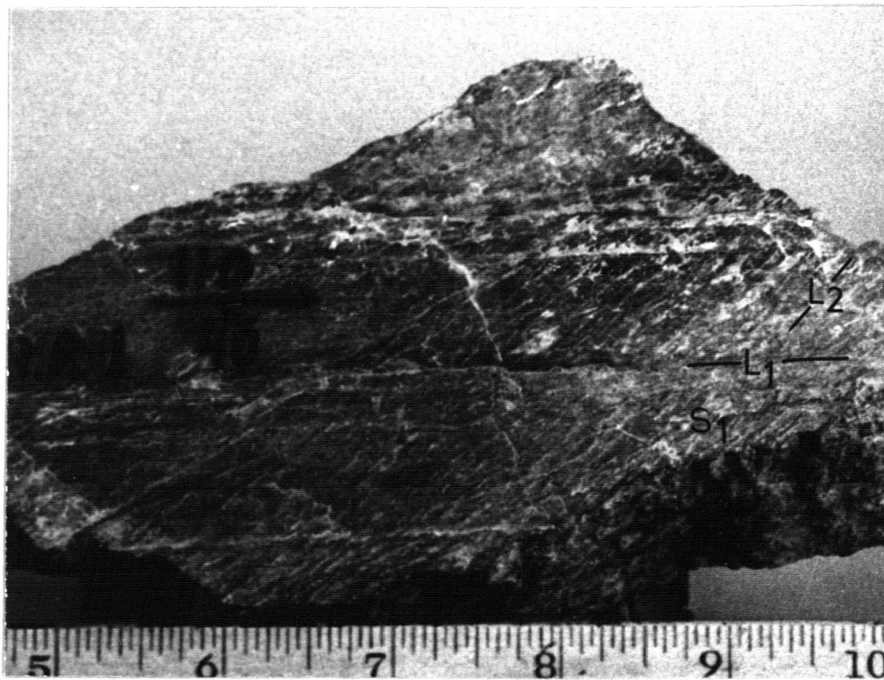
Photograph  
8 a.

Photographs 9 and 10.

Subhorizontal  $L_1$  ( $S_s/S_1$ ) lineation on  $S_1$  surface trending NW-SE, and cut across by pitching  $L_2$  ( $S_1/S_2$ ) lineation in grey slate. Locality: East of Mount Albert. Under the bridge of the old highway on the Ste Anne Northeast river.



Photograph  
9.



Photograph  
10.

Photograph 11.

Steep  $S_1$  schistosity in the Shickshock metavolcanics.

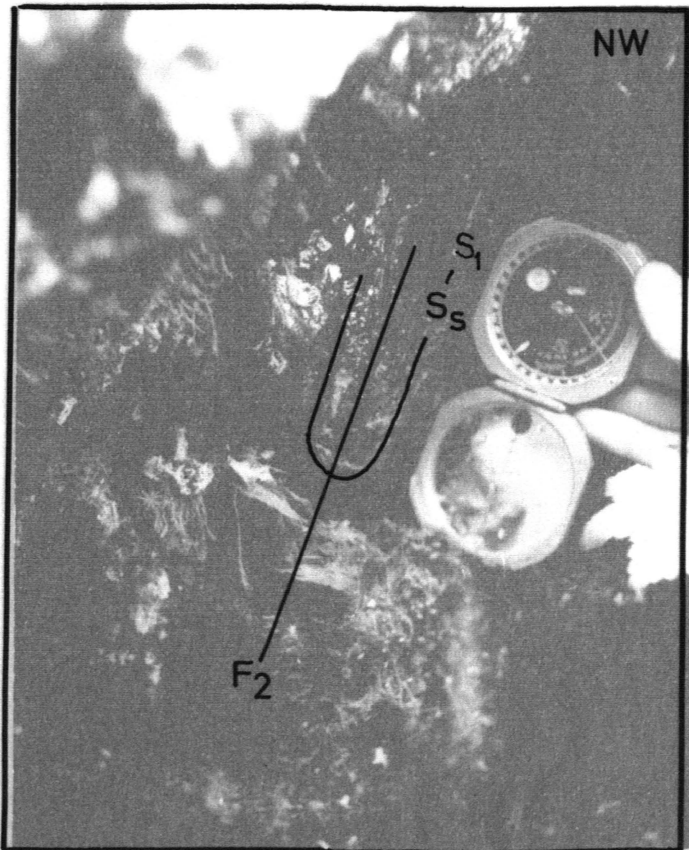
Locality: Shickshock Mountains.

Photograph 12.

Tight  $F_2$  fold indicated by an isoclinally folded secondary quartz vein. Locality: Shickshock Mountains, along Cap Chat Valley, 1 mile downstream from Wilson Brook confluence.



Photograph  
11.



Photograph  
12.



Photograph  
11.



Photograph  
12.

Photograph 13.

Sheared limestone conglomerate bed with argillaceous schistose matrix. Clasts flattened parallel to  $S_s$ .  
Locality: St.A-2/8, NE of Mount Albert.

Photograph 14.

Steeply dipping limestone beds with flat  $S_1$  schistosity crossed by  $S_2$  fracture dipping SE towards the camera, in Cambrian-Ordovician slates.  
Locality: St.A-2/11 (Appendix III), NE of Mount Albert.



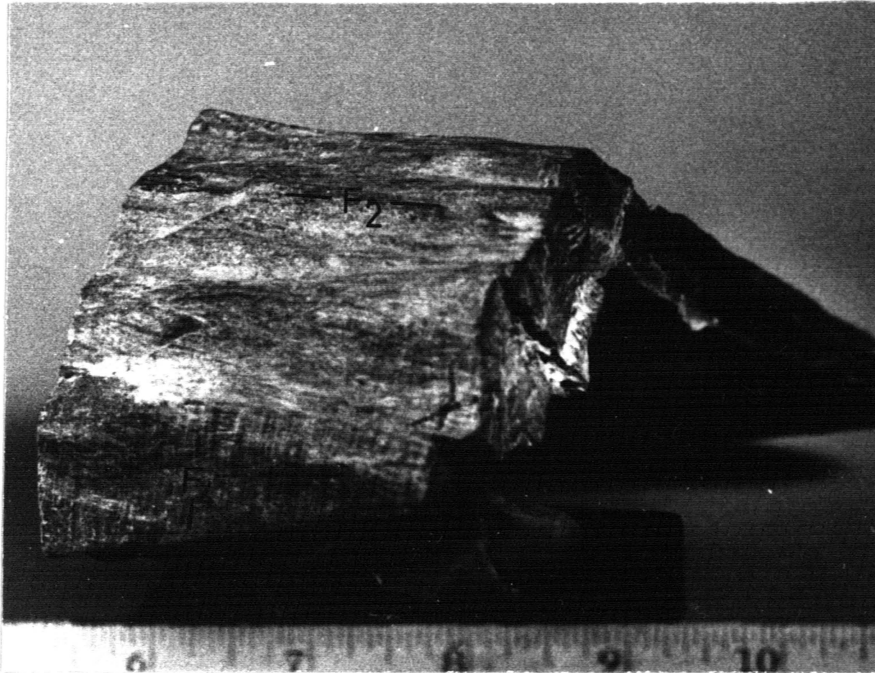
Photograph  
13.



Photograph  
14.

Photographs 15 and 16.

Steeply pitching  $L_4$  lineation on  $S_1$  schistosity,  
poorly developed along the hinge area of  $F_2$  fold  
in the Shickshock metavolcanics. Locality:  
Shickshock Mountains.



Photograph  
15.



Photograph  
16.

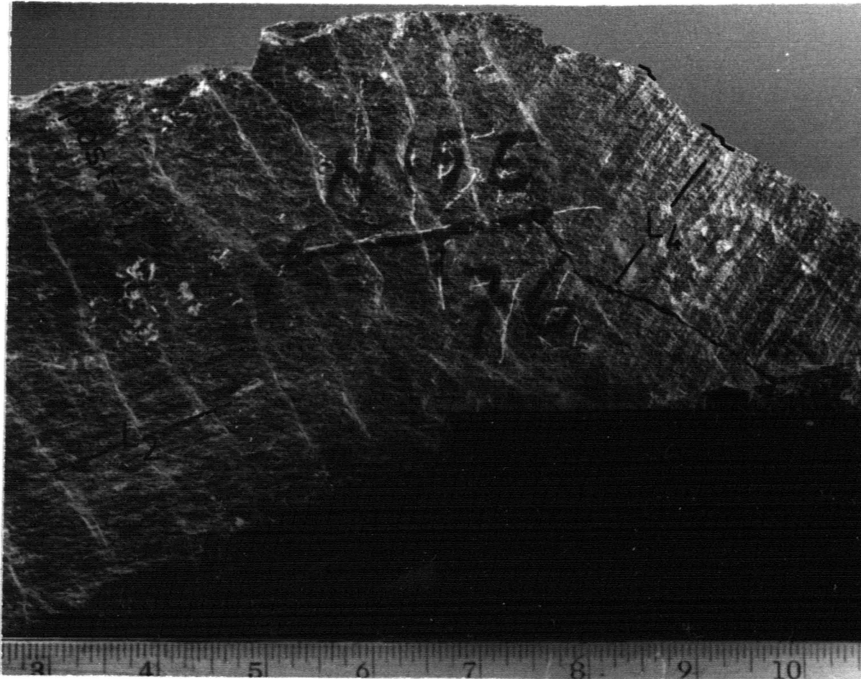
Photograph 17.

$S_1$  surface with low pitching  $L_2$  crenulation, and steeper pitching  $L_4$  (kinks) crossed by a steep probably post- $L_4$  lineation locally developed in the Shickshock metavolcanics. Locality: Shickshock Mountains.

Photograph 18.

$S_1$  surface with steeply pitching fine  $L_4$  lineation (fine kinks) crossed by a low pitching  $L_2$  crenulation. Steep Z-shaped kinks, probably post- $L_4$ , locally developed in the Shickshock metavolcanics, show an echelon shears.

Locality: Shickshock Mountains.



Photograph  
17.



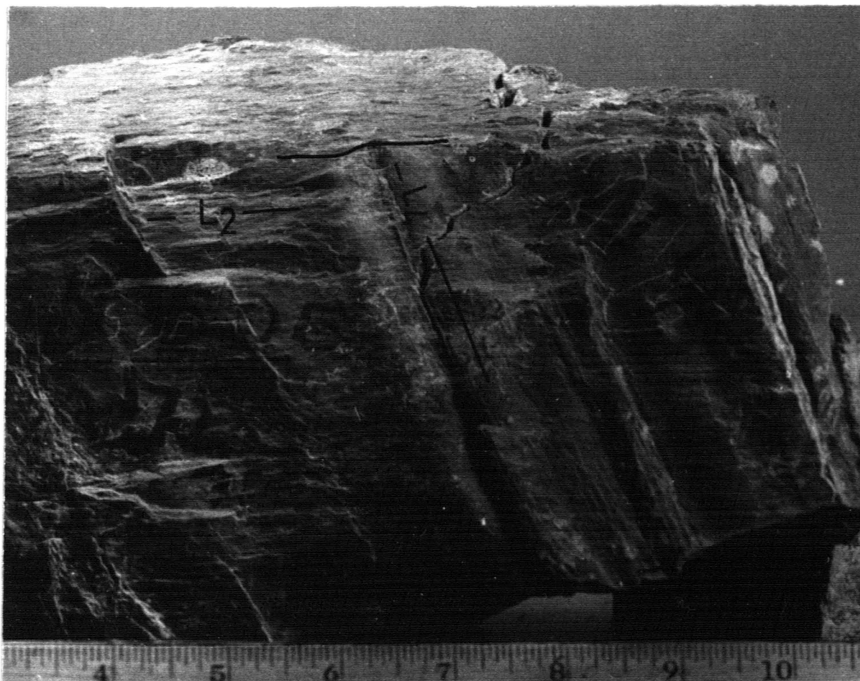
Photograph  
18.

Photograph 19.

Cambrian-Ordovician phyllitic slate containing sub-horizontal  $L_2$  crenulation on  $S_1$  schistosity surface. The steeply pitching folds with steeply pitching axial plane fracture are suspected to be  $F_4$ . Locality: MI-1/5, NE of Rene Goupil.

Photograph 20.

Cambrian-Ordovician phyllitic grey slate with steep  $L_4$  lineation (kink) on  $S_1$  surface.  $L_2$  is a low pitching lineation on the  $S_1$  surface. Locality: 4 miles NE of Sayabec village (AMQ-10/3).



Photograph  
19.



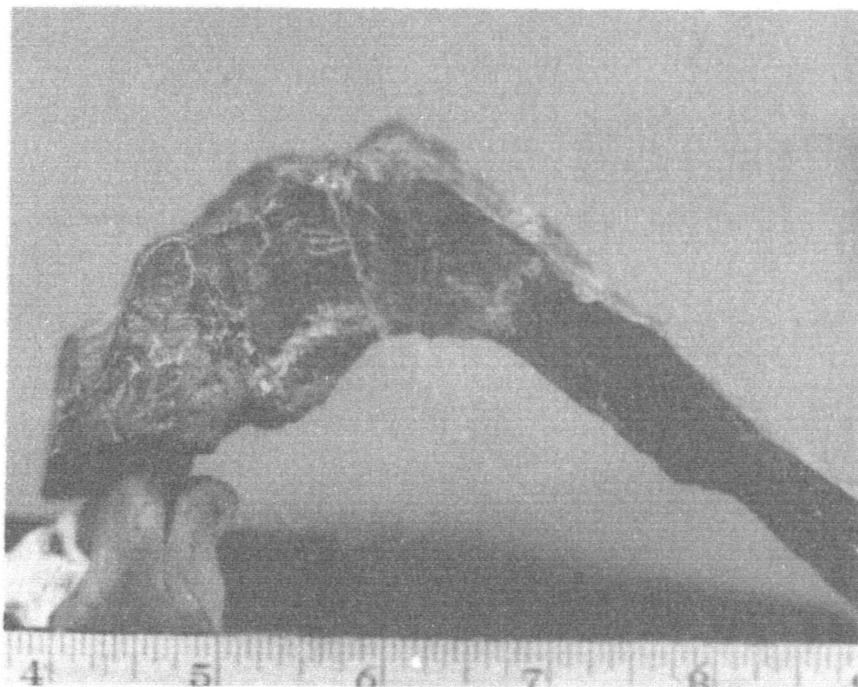
Photograph  
20.

Photograph 21.

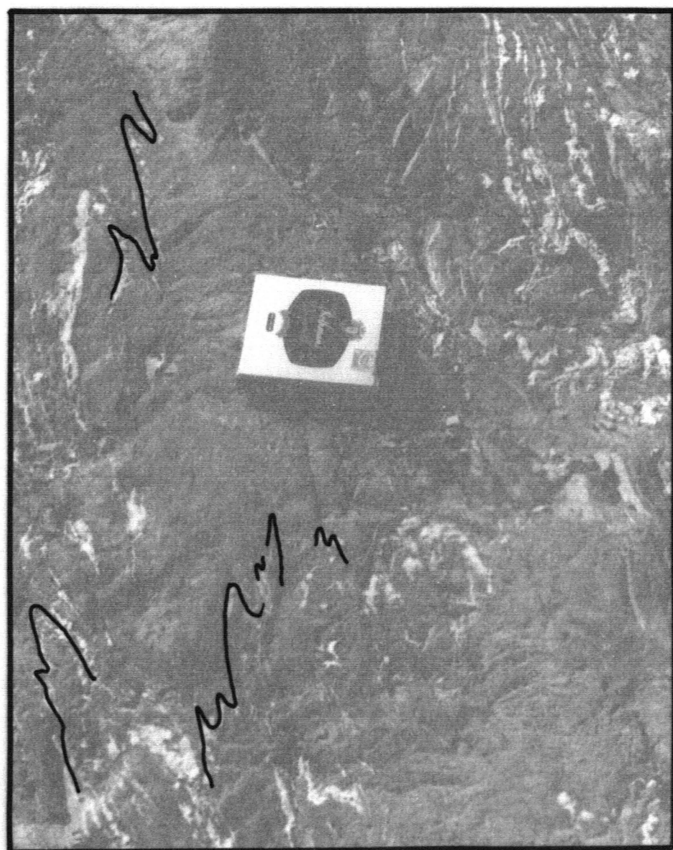
$F_2$  fold deforming the  $S_1$  schistosity in the Shickshock metavolcanics.  $F_2$  fold demonstrates thickening in hinge area. Locality: SK-3, west bank of Matane Lake in the Shickshock Mountains.

Photograph 22.

Intricate  $F_2$  folding of secondary quartz veins in the Shickshock metavolcanics. Locality: Shickshock Mountains.



Photograph  
21.



Photograph  
22.



Photograph  
21.



Photograph  
22.

Photographs 23 and 24.

Open  $F_2$  flexures in calcareous slates of Quebec Group. Note the poor development of the  $S_2$  cleavage in the hinge area. Locality: ME-11, St. Lawrence coast.



Photograph  
23.



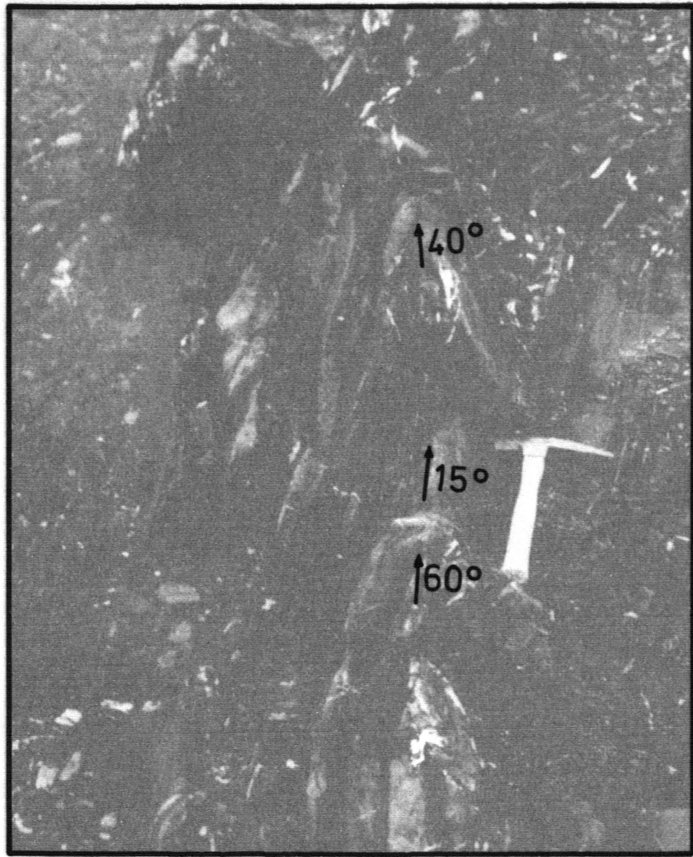
Photograph  
24.

Photograph 25.

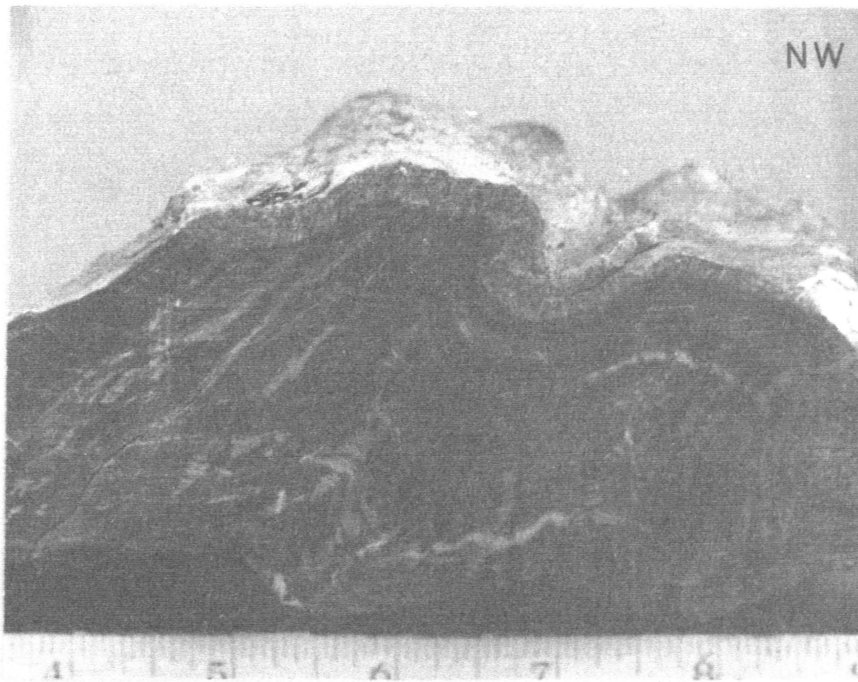
F<sub>2</sub> fold in grey calcareous slate of Quebec Group demonstrating variation in plunge. Locality: ½ mile SW of ME-1, along St. Lawrence coast.

Photograph 26.

F<sub>2</sub> folds in calcareous slates of the Quebec Group. Calcareous enrichment along the axial planes is secondary and related to the development of folds. Locality: Near Riviere Matane.



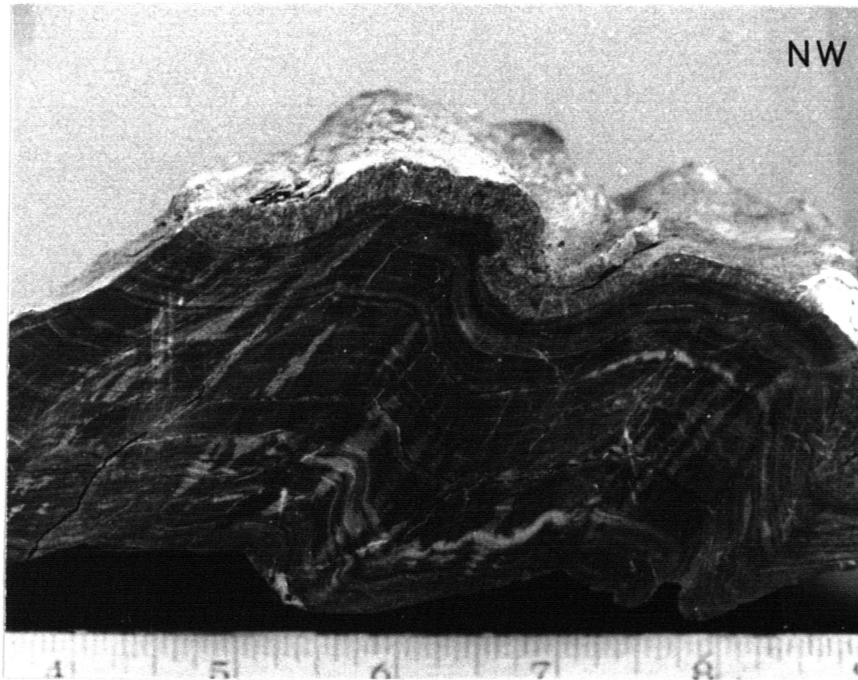
Photograph  
25.



Photograph  
26.



Photograph  
25.



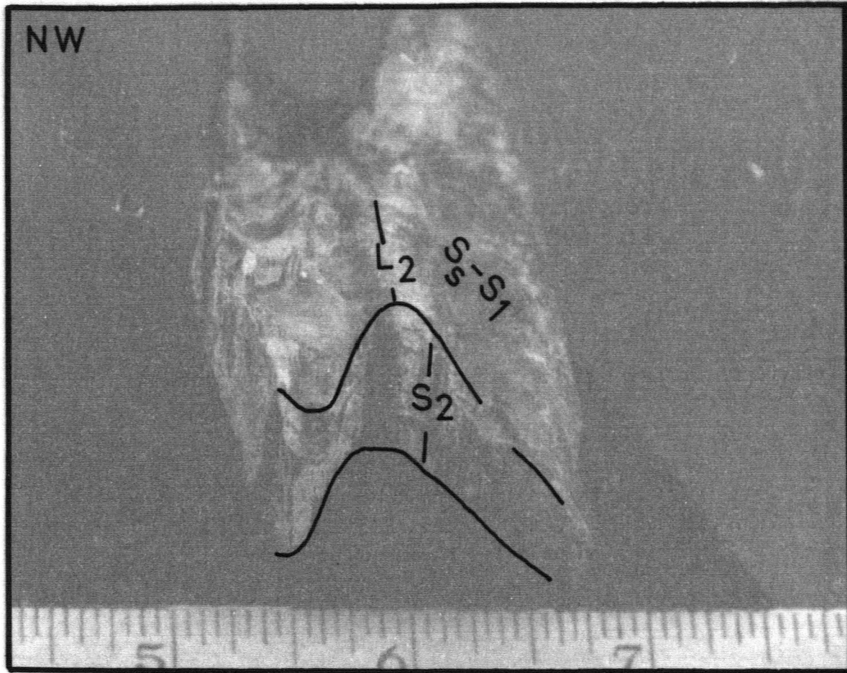
Photograph  
26.

Photograph 27.

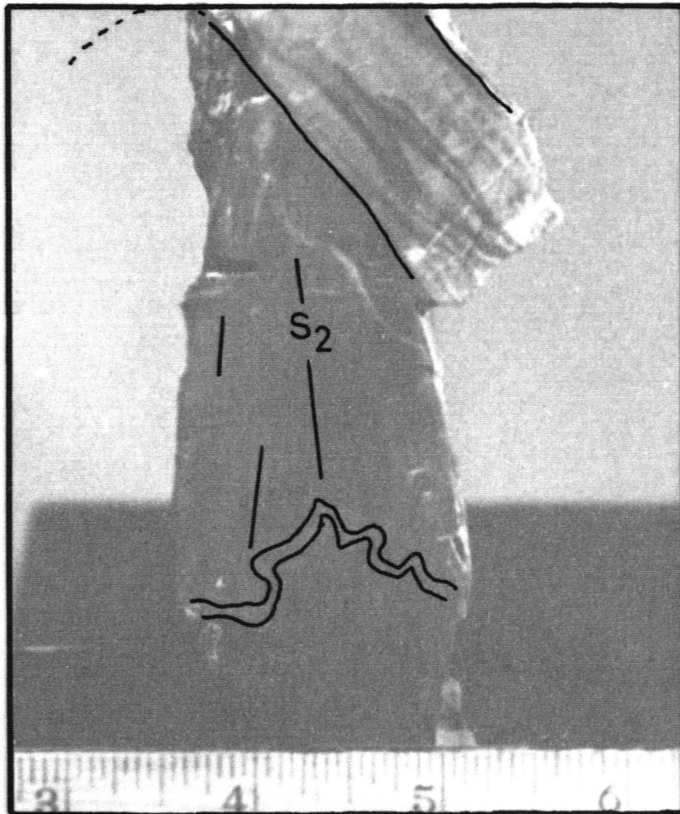
$F_2$  fold deforming  $S_1$  schistosity parallel to bedding in Cambrian-Ordovician grey slate. Note the irregular hinge of the fold on  $S_s$  and  $S_1$  surface. Locality: MI-3/1, 6 miles NE of Riviere Matane village.

Photograph 28.

Difference in  $F_2$  fold style between two beds of calcareous slate of different thicknesses. Note also the  $S_2$  cleavage in the Ordovician slate. Locality: St. Lawrence coast.



Photograph  
27.



Photograph  
28.



Photograph  
27.



Photograph  
28.

Photograph 29.

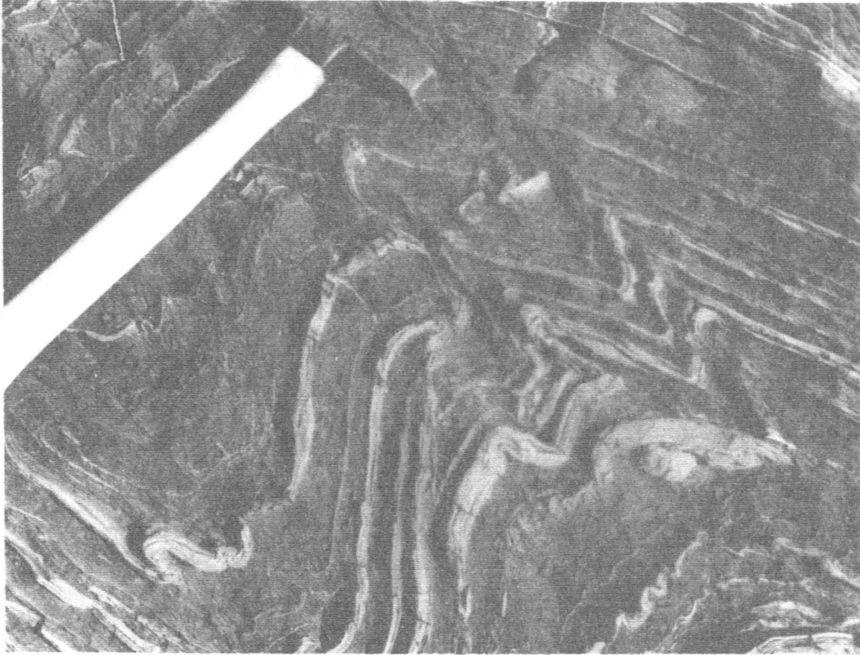
F<sub>2</sub> fold in calcareous slate. Locality: ME-1,  
St. Lawrence coast.



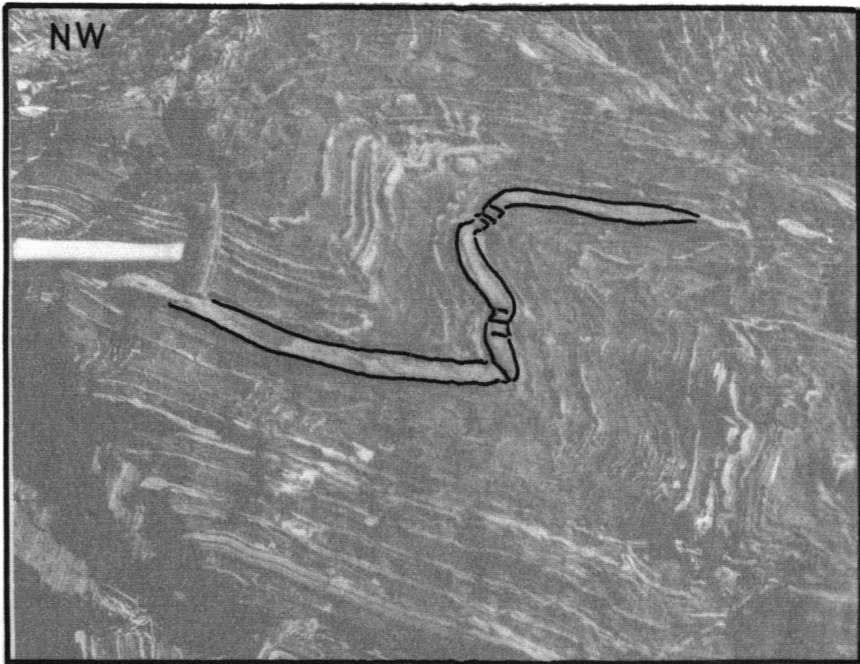
Photograph  
29.

Photographs 30 and 31.

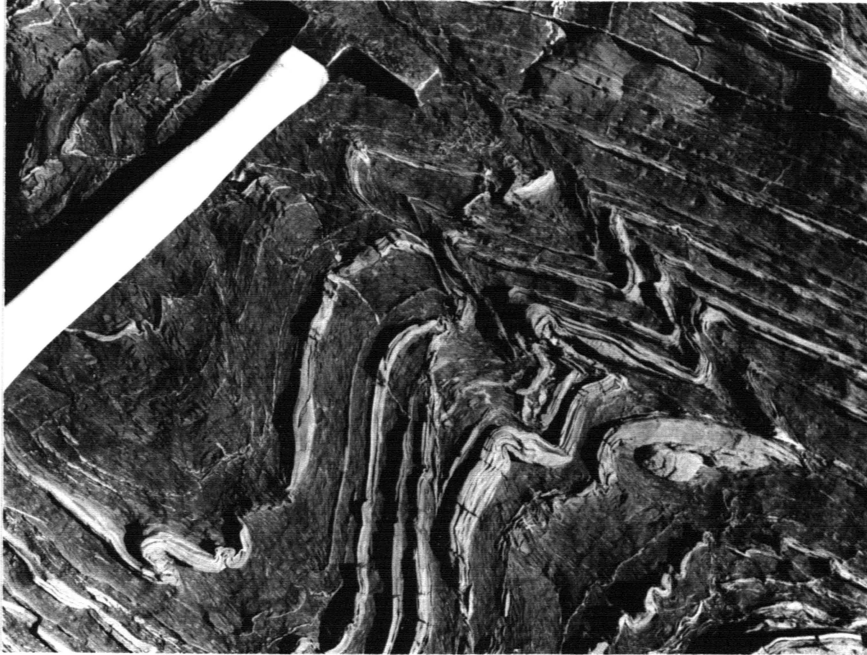
F<sub>2</sub> folds in laminated slate. The competent calcareous siltstone layers are thinner in the limb areas, also occasionally in the hinge areas, the competent layers are broken apart and 'intruded' by slate. Note also the disharmony in folds. Locality: ME-1, St. Lawrence coast.



Photograph  
30.



Photograph  
31.



Photograph  
30.



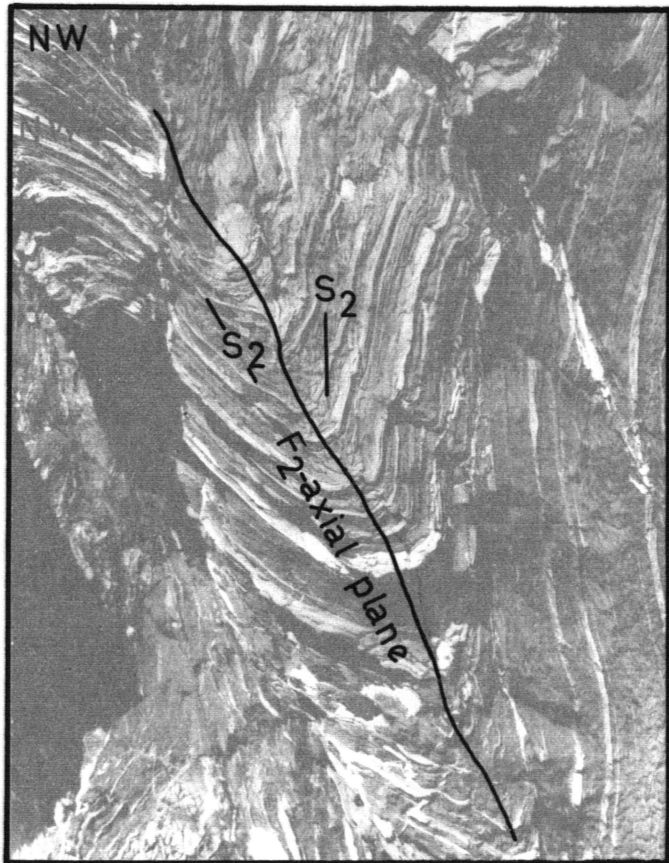
Photograph  
31.

Photograph 32.

$F_2$  folds in laminated slates of Quebec Group.

$S_2$  cleavage diverges upward in the syncline.

Locality: ME-1, St. Lawrence coast.



Photograph  
32.



Photograph  
32.

Photographs 33 and 34.

Plan sections of  $F_2$  folds in slate with calcareous siltstone layers in the St. Lawrence coastal area. The coastal area during low tides show some of the best outcrops for the study. Locality: 14 miles SW of Matane.



Photograph  
33.



Photograph  
34.

Photograph 35.

F<sub>2</sub> fold in Ordovician slate. The competent siltstone layers are disharmonically folded. The overlapping of the siltstone layer in the left central part (due to a contraction fault) reflects shortening across the fold.

Locality:  $\frac{1}{2}$  mile SW of Grosses Roches.

St. Lawrence coast.

Photograph 36.

Isoclinal tight F<sub>2</sub> fold in the slates of the Quebec Group. Locality:  $\frac{1}{2}$  mile SW of Grosses Roches, St. Lawrence coast.



Photograph  
35.



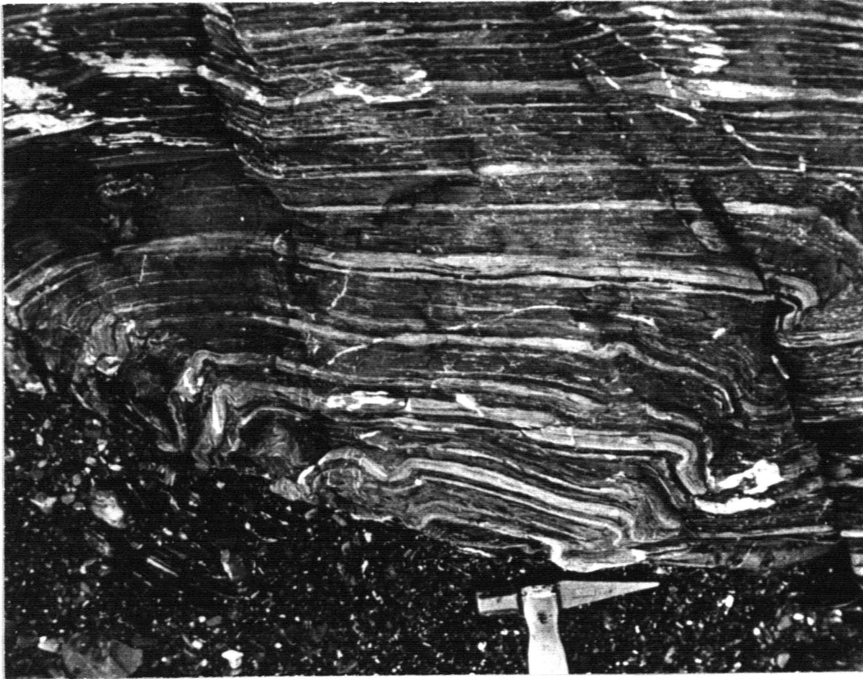
Photograph  
36.

Photograph 37.

Conjugate  $F_2$  folds in the calcareous slate of the Quebec Group. Locality: ME-1, St. Lawrence coast.

Photograph 38.

$F_2$  fold in interbedded slate and siltstone of the Quebec Group. The siltstone beds in the hinge area are crossed by tensional joint fractures. Locality: ME-1, St. Lawrence coast.



Photograph  
37.



Photograph  
38.

Photograph 39.

Disharmonic  $F_2$  folds in laminated slate.

Locality: ME-4, St. Lawrence coast.

Photograph 40.

Disharmonic  $F_2$  folds in laminated slate of

Quebec Group. Locality: MW-8,

St. Lawrence coast.



Photograph  
39.



Photograph  
40.

Photograph 41.

F<sub>2</sub> fold in the Quebec Group. The calcareous layers show a thinning along the fold limbs. Note a lack of cleavage along the hinge area of the folds. Locality: ME-6, St. Lawrence coast.

Photograph 42.

Concentric F<sub>2</sub> fold in calcareous slate in the Quebec Group. The outer part of competent layers in the fold are crossed by tensional fractures. Locality: ME-6, St. Lawrence coast.



Photograph  
41.



Photograph  
42.

Photograph 43.

Isoclinal  $F_2$  folds in calcareous slate, Quebec Group. Locality: ME-1, St. Lawrence coast.

Photograph 44.

A typical planed  $F_2$  fold outcrop in calcareous slate with siltstone layers in the St. Lawrence coastal area.  $S_2$  reticulate cleavage converges upwards in the anticline. The thinning of the competent layer is probably sedimentary as lithology of the layer changes across the fold. Locality: ME-1, St. Lawrence coast. See plate IIc5.



Photograph  
43.



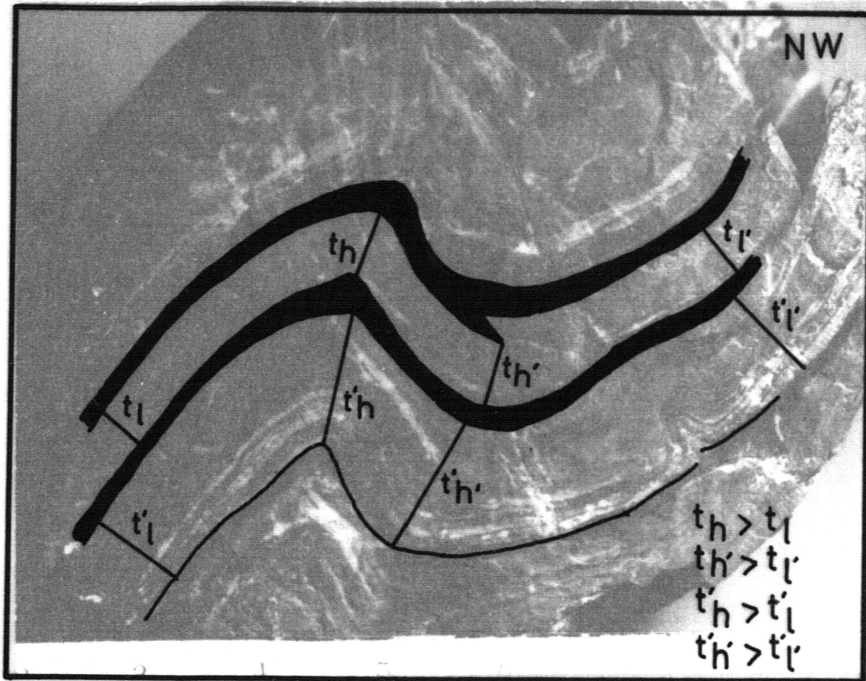
Photograph  
44.

Photograph 45.

Calcareous siltstone with minor shale intercalations in Quebec Group. Thickening in both siltstone and shale occurs along the hinges of the folds. The shale layers contain bedding foliation with slickensides parallel to the dip. Locality: MW-2, St. Lawrence coast.

Photograph 46.

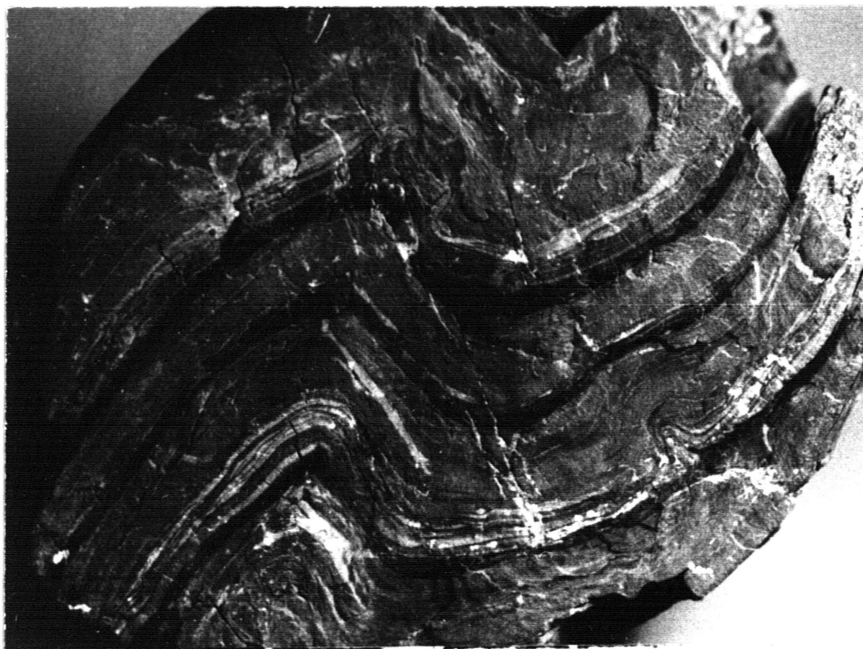
F<sub>2</sub> folds in the hinge area of a plunging regional fold. S<sub>2</sub> fracture cleavage is present in the hinge area. Locality: St. Lawrence coast.



Photograph  
45.

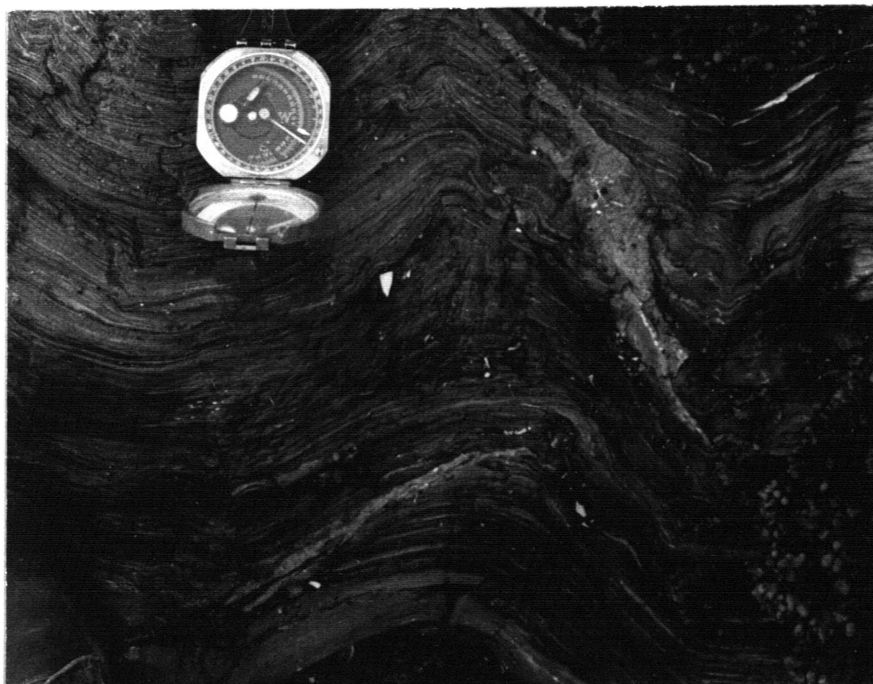


Photograph  
46.



Photograph  
45.

0 2 4 5 6



Photograph  
46.

Photograph 47.

Double plunging canoe shaped  $F_2$  folds in calcareous slate of Quebec Group. Locality: St. Lawrence coast, MW-5/6.

Photograph 48.

Exposure of the Quebec group slates at low tide.  
Locality: ME-11, St. Lawrence coastal area.



Photograph  
47.



Photograph  
48.

Photograph 49.

Finely bedded slates tightly plicated by  $F_2$  folds enclosed in more openly folded competent layers in the Quebec Group. Locality: St. Lawrence coast.

Photograph 50.

Open  $F_2$  fold in calcareous slates in the Fortin slate group, Devonian. Steep  $S_2$  cleavage is weakly developed. Locality: 18 miles NW of Matapedia town.



Photograph  
49.



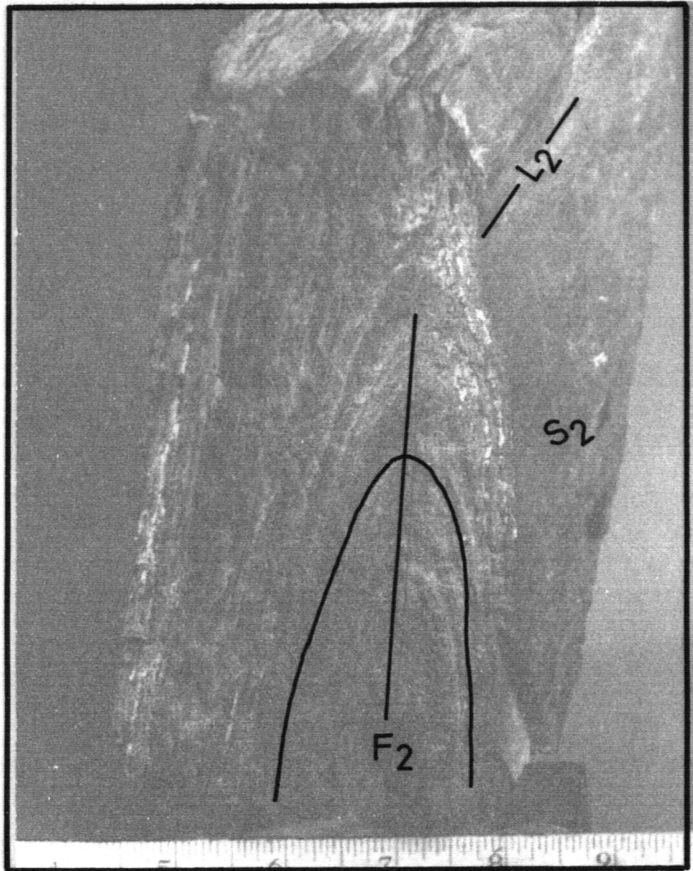
Photograph  
50.

Photograph 51.

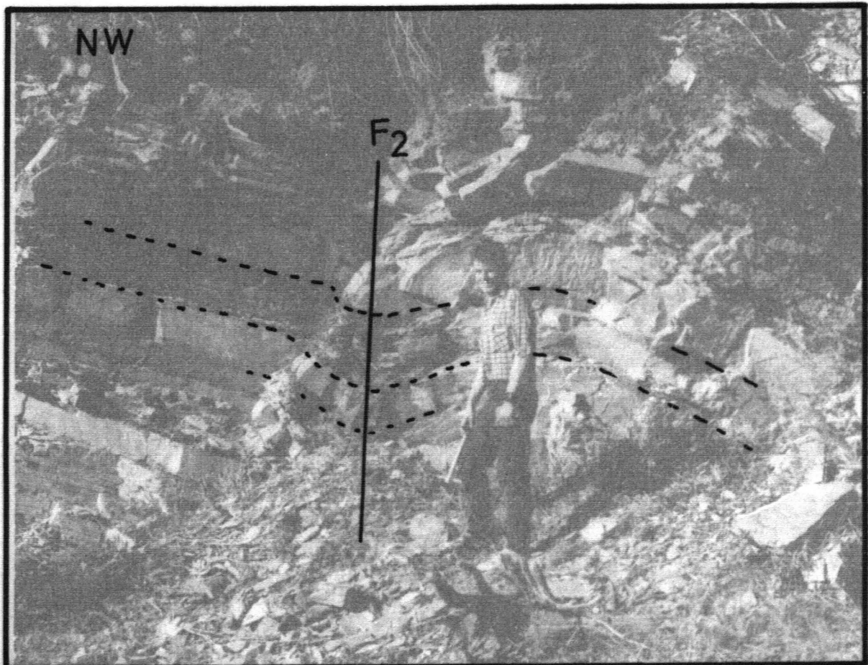
Tight 'similar' type  $F_2$  fold in sericitic slate Fortin Group, Devonian. The  $S_2$  cleavage is well developed. Locality:  $\frac{1}{2}$  mile east of Ste Florence village.

Photograph 52.

Open monoclinal  $F_2$  fold typical of St. Leon Formation 4 miles E of Matane Lake, and south of the Shickshock Mountains.



Photograph  
51.



Photograph  
52.



Photograph  
51.



Photograph  
52.

Photograph 53.

S<sub>2</sub> cleavage in a F<sub>2</sub> syncline diverging upwards in Quebec Group slates. Locality: MW-5/6, St. Lawrence coast.

Photograph 54.

Upward divergent S<sub>2</sub> fracture cleavage in an F<sub>2</sub> syncline. Note the competent siltstone layers contraction faulted, and also punched outward from the core of the fold. (The camera lens cap is 1½" in diameter). Locality: St. Lawrence coastal area, MW-5/6.



Photograph  
53.



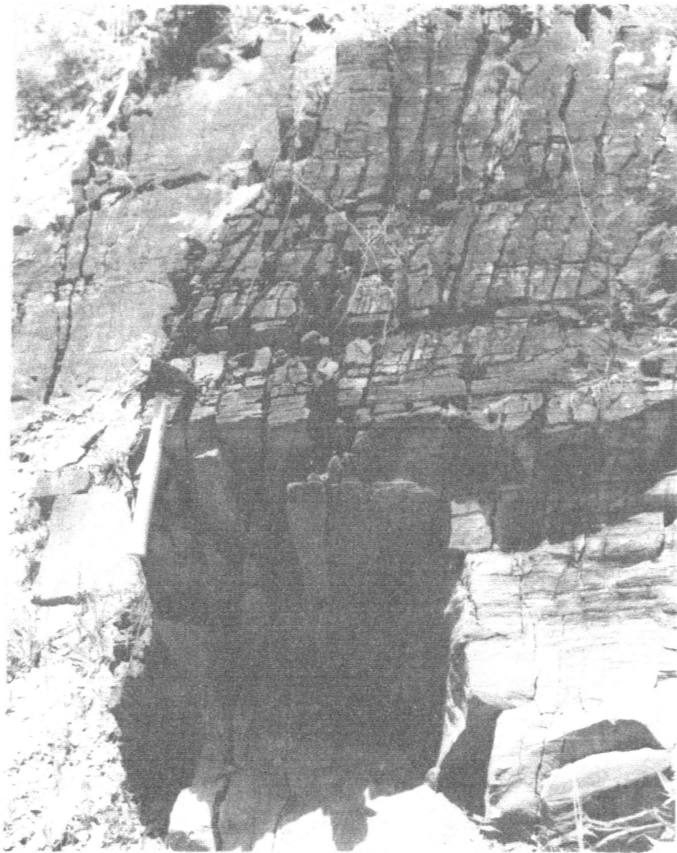
Photograph  
54.

Photograph 55.

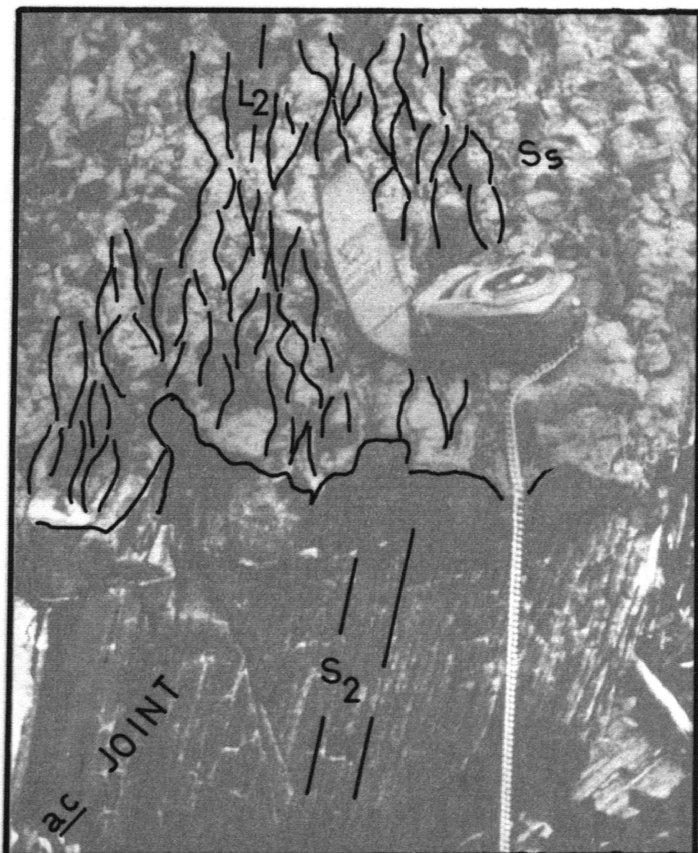
Vertical  $S_2$  fractures in Grande Greve siltstone, Devonian. The beds dip NE. The  $L_2$  lineation ( $S_s/S_2$ ) suggests steep NE plunge of the anticlinal structure. Locality: 2 miles SW of Casault Lake.

Photograph 56.

Recticulated  $S_2$  fracture cleavage trace on low dipping bedding in calcareous slate of Quebec Group. Bulges of irregular persistence along the  $S_s/S_2$  intersections occur between the  $S_2$  surfaces. They are regarded as due to a shortening between adjacent  $S_2$  planes. Locality: MW-5/6, St. Lawrence coast.



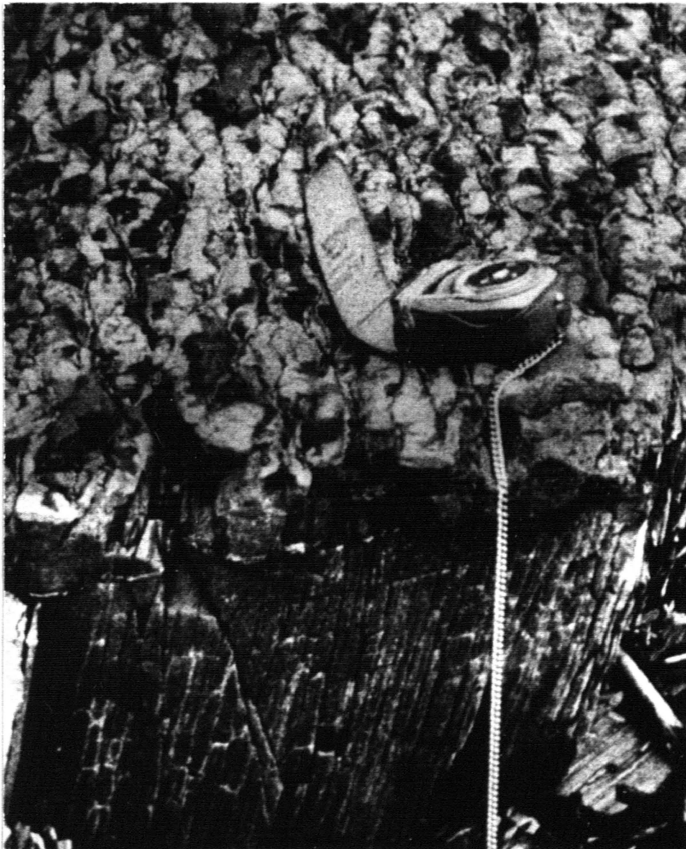
Photograph  
55.



Photograph  
56.



Photograph  
55.



Photograph  
56.

Photograph 57.

Plastically deformed microlithon along the hinge of a  $F_2$  fold in Fortin Slates.

Locality: 1 mile south of Routhierville, near Ste. Florence.

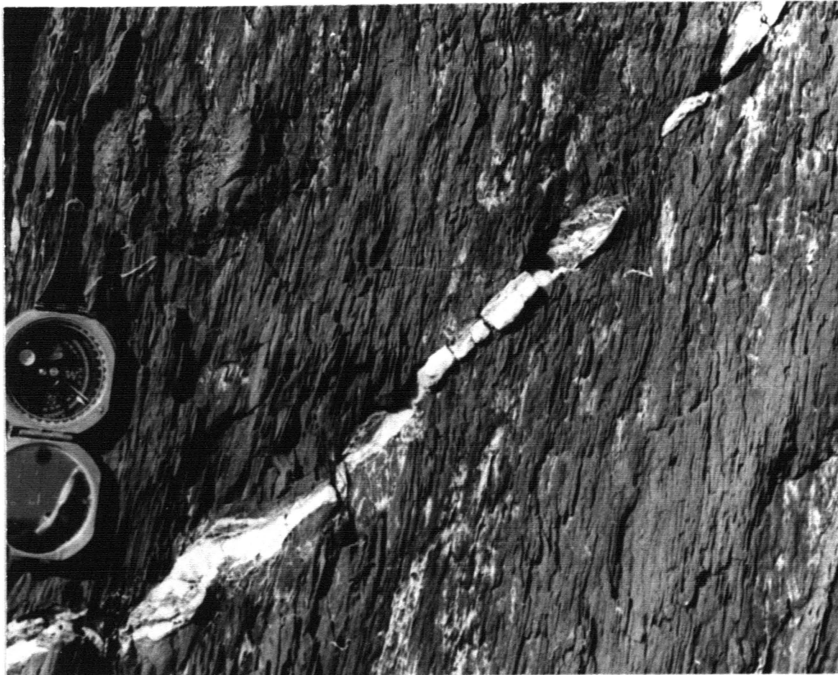
Photograph 58.

Calcareous siltstone layer broken into discontinuous lenses and crossed by  $S_2$  cleavage.

The cleaved rock is calcareous slate of Quebec Group. Locality: ME 1, St. Lawrence coast.



Photograph  
57.



Photograph  
58.

Photograph 59.

Fortin Slate closely fractured by  $S_2$  cleavage.

Note the banding due to colour variation of low dipping beds on ac joint surface.

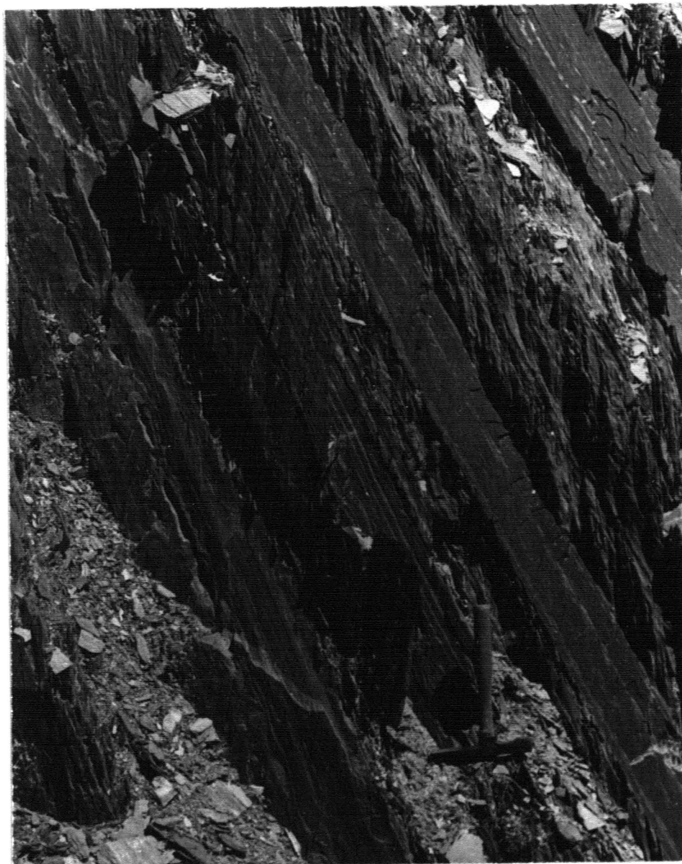
Locality: Near Causapscal.

Photograph 60.

$S_2$  fracture cleavage in Fortin Slate. Calcareous layers resistant to weathering, stand out as poorly cleaved bands, whereas in non-calcareous layers the cleavage is more obvious due to deeper weathering. Locality: Near Causapscal.



Photograph  
59.



Photograph  
60.

Photograph 61.

Finely developed  $S_2$  fracture cleavage in argillaceous Cape Bon Ami Formation. Boudin like objects are calcite rich bodies parallel to bedding. Locality: 2 miles SW of Casault Lake.

Photograph 62.

Cleavage in Cape Bon Ami Formation. Locality: Along Truite River,  $\frac{1}{2}$  mile upstream of Matane River confluence.



Photograph  
61.



Photograph  
62.

Photograph 63.

Coarsely developed  $S_2$  fracture cleavage in  
openly folded York River-Heppel Formations.  
Locality: North of Casault Lake.

Photograph 64.

Coarsely developed  $S_2$  fracture cleavage in  
coarse sandstone of York River-Heppel Forma-  
tions. Locality: North of Casault Lake.



Photograph  
63.



Photograph  
64.

Photographs 65 and 66.

Fractures in calcareous slate of Quebec Group between the  $S_2$  fracture cleavage planes.

Photo. 66 is the same fracture as in Photo.65, as seen on dip slope of the cleavage. The fractures are normal to  $S_2$ , which they meet but do not cross and form irregular polygons between the  $S_2$  cleavage planes. (Their origin is unaccounted for, and may have formed by a contraction phenomenon similar to that in mud-cracks). Locality: North of Mount Joli along the St. Lawrence coast.



Photograph  
65.



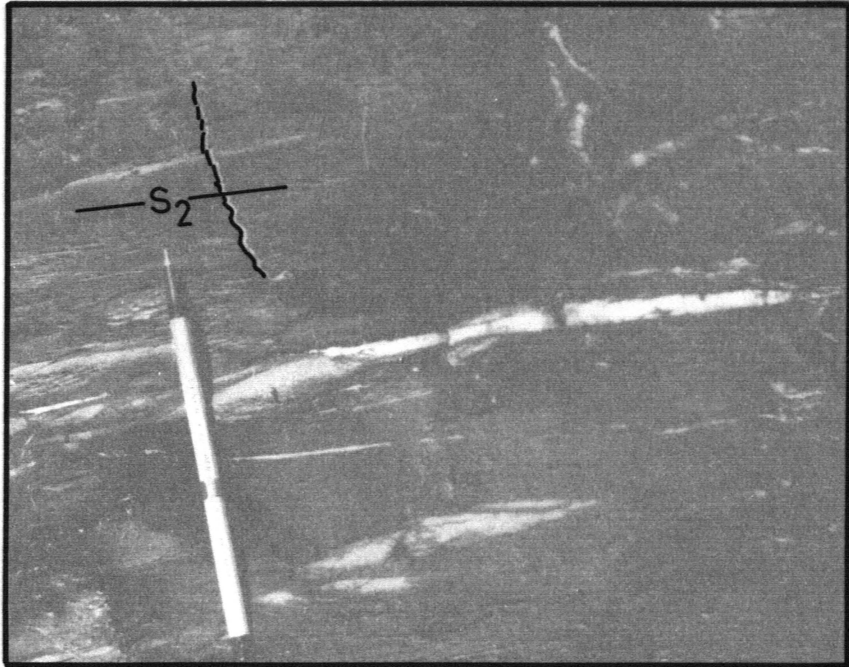
Photograph  
66.

Photograph 67.

Fine calcite veins cutting across  $S_2$  cleavage in Quebec Group slate. The cross calcite veins are displaced and folded across cleavage. Locality: West of Baie des Sables.

Photograph 68.

A post- $F_2$  calcite vein containing a lineation caused by the intersection of the layering and calcite vein. Locality: SW of Baie des Sables.



Photograph  
67.



Photograph  
68.



Photograph  
67.



Photograph  
68.

Photograph 69.

Calcite vein oblique to layering but sub-parallel to cleavage in Quebec Group slate.

Locality: MW-9, St. Lawrence coast.

Photograph 70.

Calcite vein parallel to cleavage older than the vein at high angle parallel to ac joint.

Locality: St. Lawrence coast.



Photograph  
69.



Photograph  
70.

Photograph 71.

Calcite vein sub-parallel to the  $S_2$  fracture cleavage. Locality: MW-7, St. Lawrence coast.

Photograph 72.

Steeply plunging slickensides on calcite vein in Quebec Group slate. Locality: MW-7, St. Lawrence coast.



Photograph  
71.



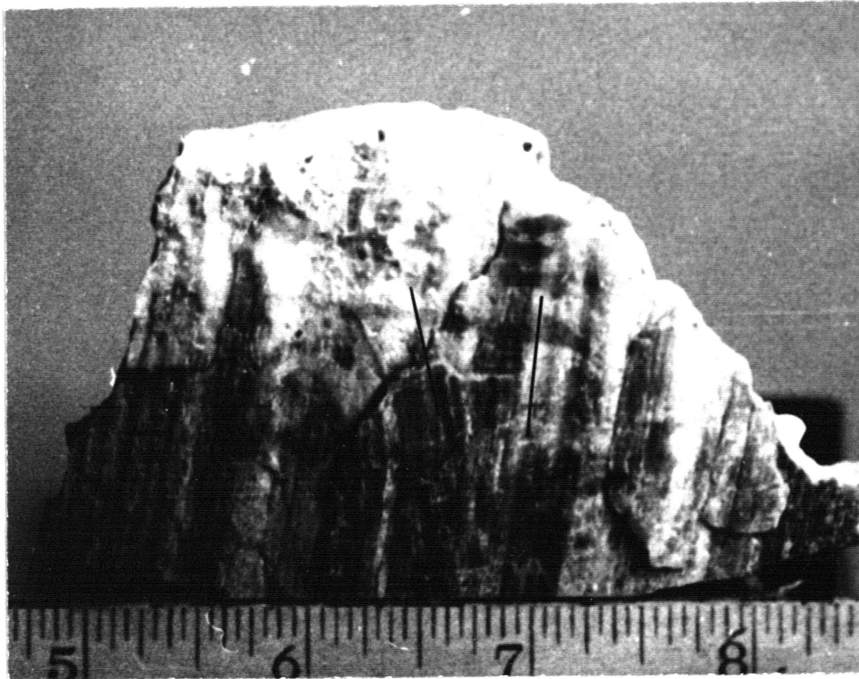
Photograph  
72.

Photograph 73.

Steeply plunging slickensides accompanied by diagonally plunging set. Locality: 1 mile SW of St. Ulric.

Photograph 74.

Diagonally pitching slickensides on calcite vein oblique to  $S_2$  in grey slates of Quebec Group. Locality: 1 mile SW of St. Ulric.



Photograph  
73.



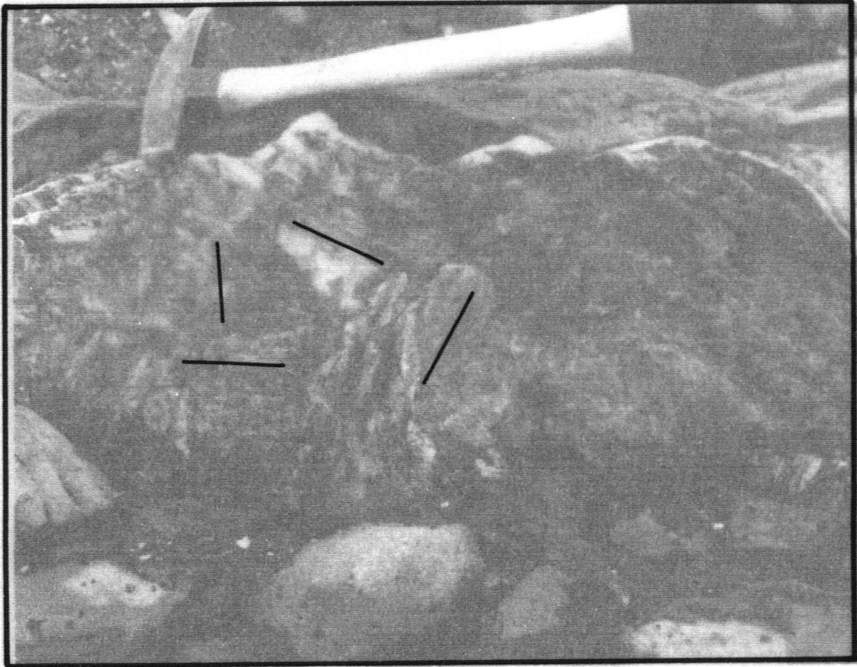
Photograph  
74.

Photograph 75.

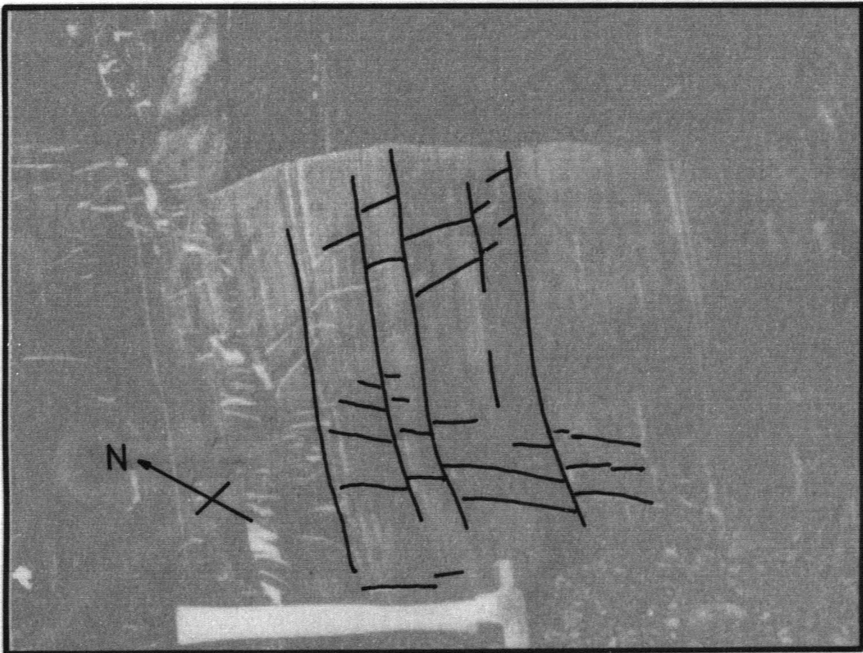
Calcite vein surface containing multiple (at least 4) sets of slickensides in grey slates of Quebec Group. Locality: SW of St. Ulric.

Photograph 76.

Plan view of calcite veins offset sinistrally across  $S_2$  cleavage planes. The two sinistrally displaced veins are inclined to each other. The rock type is grey slates of Quebec Group. Locality: 5 miles NE of Ste. Felicite, St. Lawrence coast.



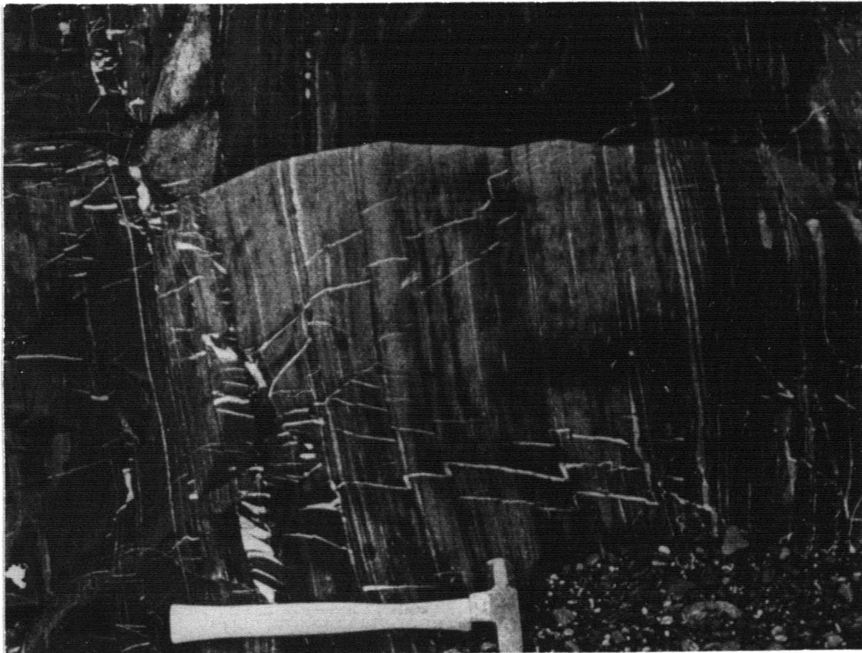
Photograph  
75.



Photograph  
76.



Photograph  
75.



Photograph  
76.

Photograph 77.

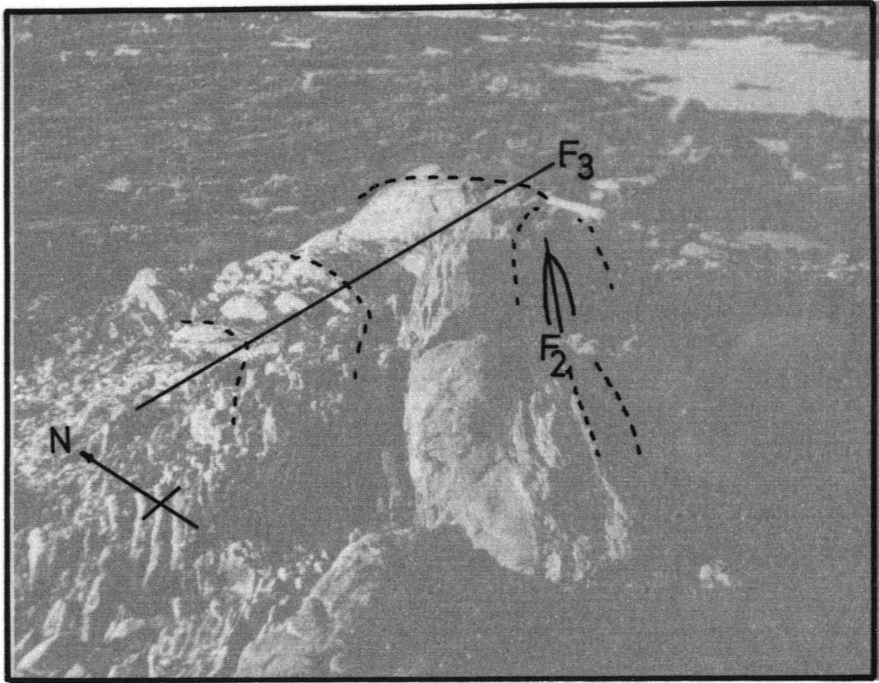
$F_2$  fold deformed by a s-shaped  $F_3$  cross fold  
in the calcareous slate in Quebec Group.

Locality: ME 11a, SW of Ste. Felicite.

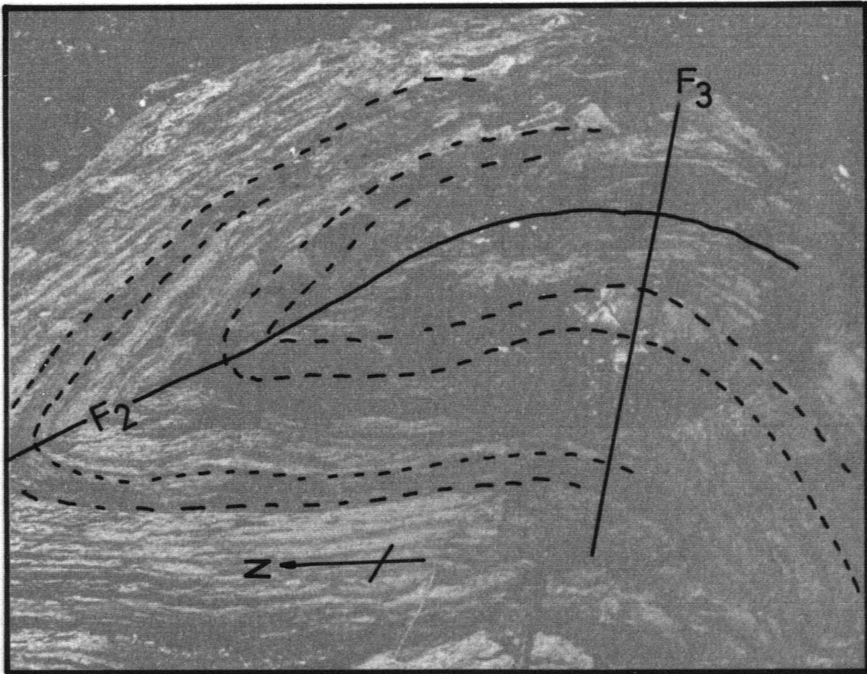
Photograph 78.

$F_2$  fold deformed by a S-shaped  $F_3$  cross fold.

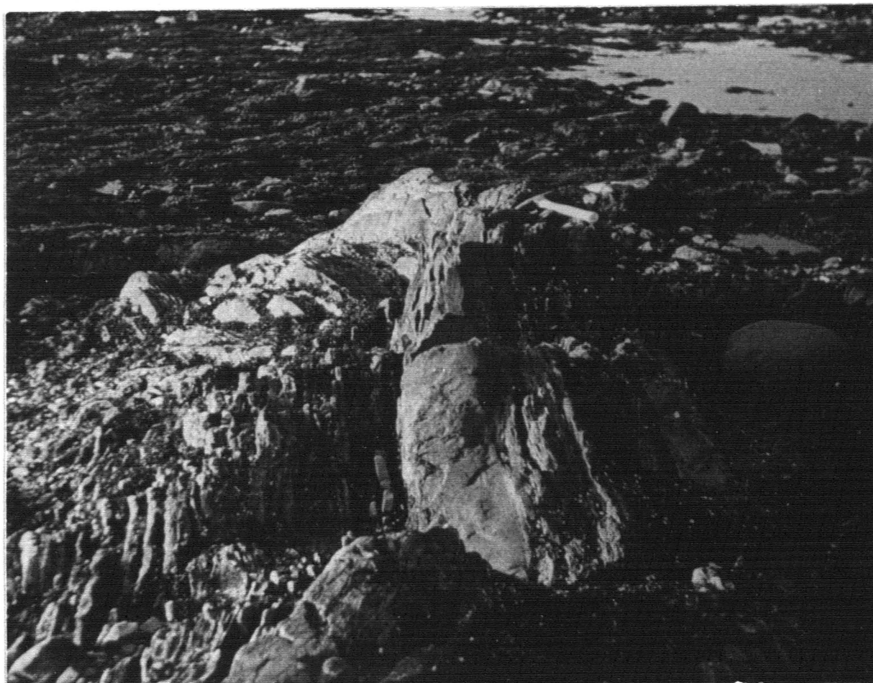
Locality: MW-9, 1 mile SW of St. Ulric.



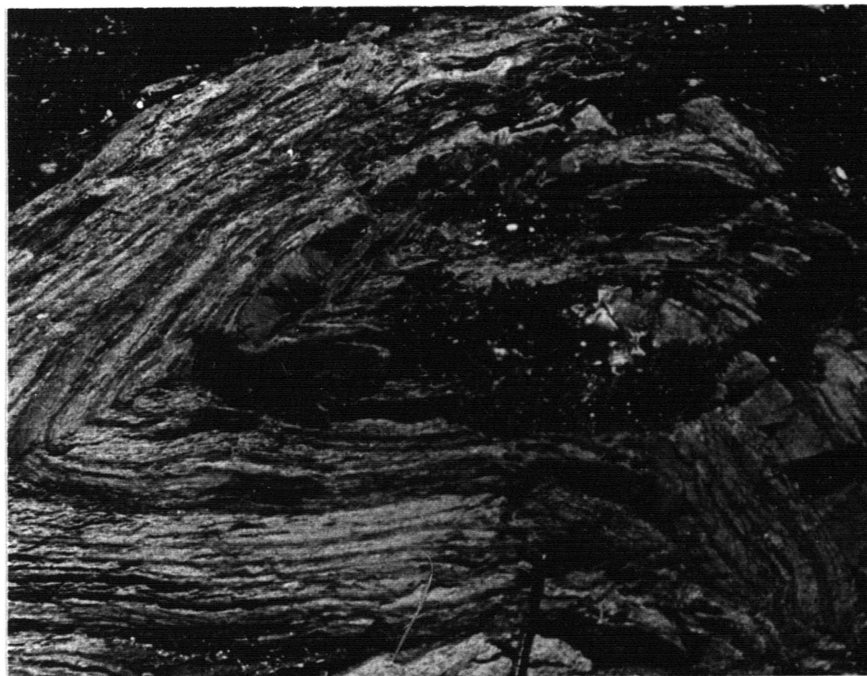
Photograph  
77.



Photograph  
78.



Photograph  
77.



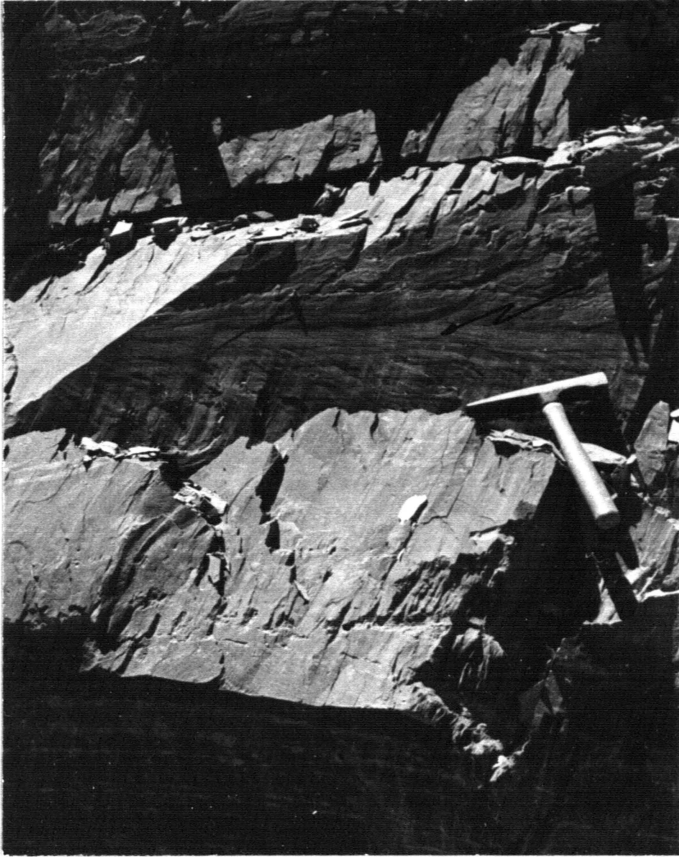
Photograph  
78.

Photograph 79.

Plumose structure on dip joint surface in the Fortin Slate of Devonian age. Note well developed  $S_2$  cleavage. Locality: Near Causapscal township.

Photograph 80.

Plumose structure on strike joint surface in the impure grey sandstone of Quebec Group. Locality: ME 7, St. Lawrence coast.



Photograph  
79.



Photograph  
80.

Photograph 81.

Multiple dip and strike joint fractures.

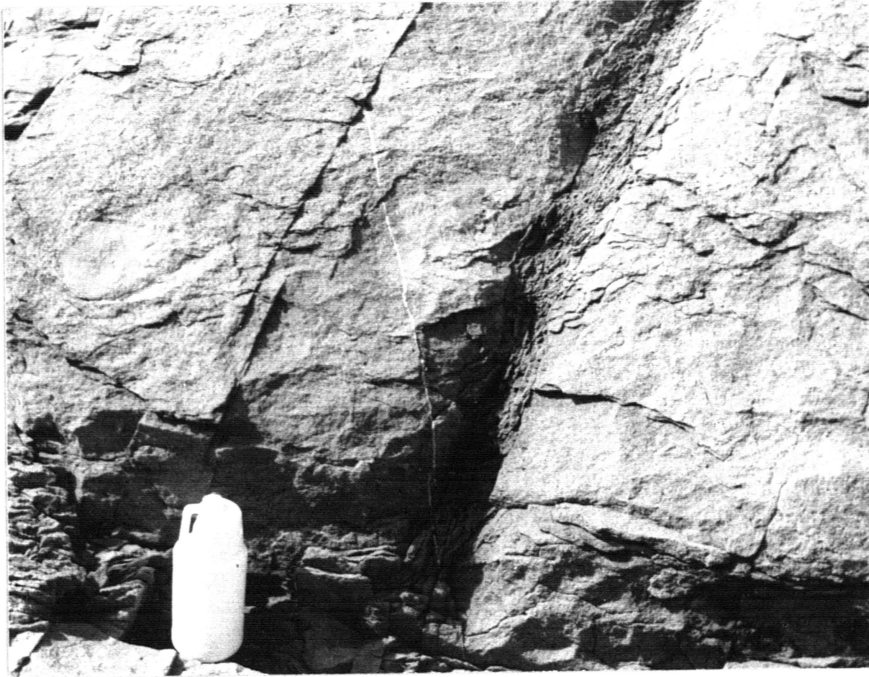
The joint 'zones' are deeply weathered.

Locality: ME 7, St. Lawrence coast.

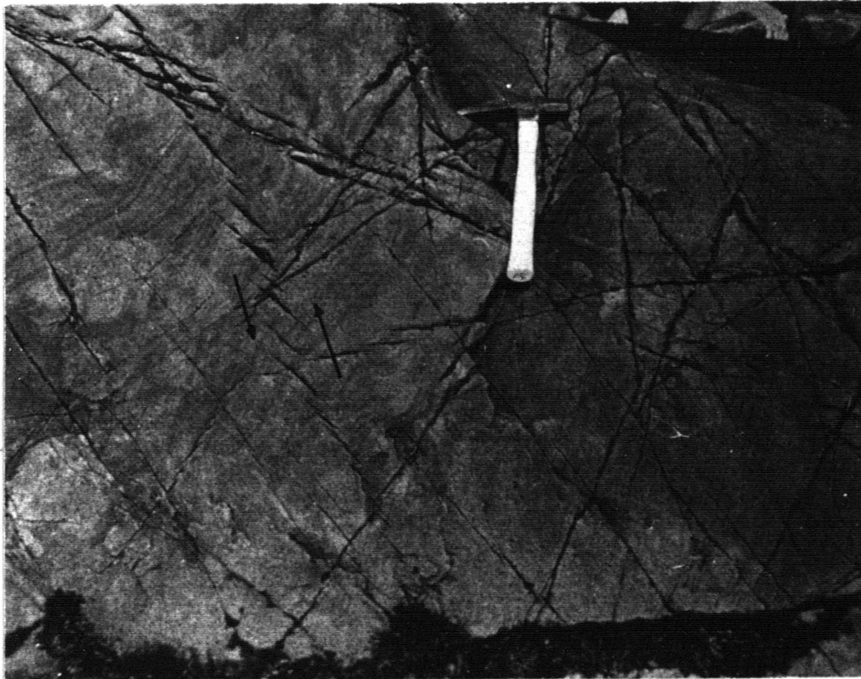
Photograph 82.

Oblique joint sets formed en echelon,  $15^{\circ}$   
to joint trace, on the dip surface of the  
impure grey sandstone of Quebec Group.

Locality: ME-7, St. Lawrence coast.



Photograph  
81.



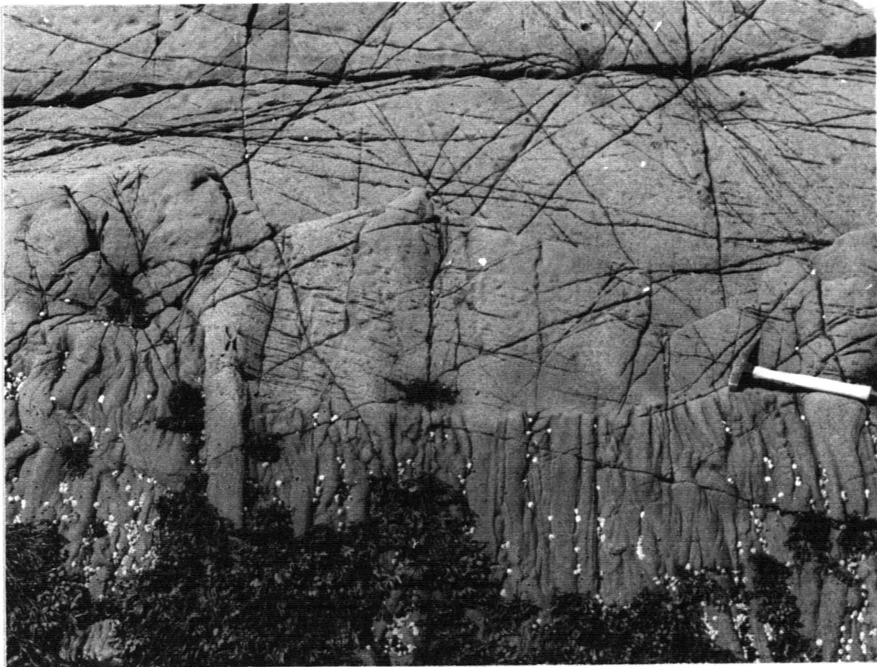
Photograph  
82.

Photograph 83.

Inverted well jointed impure grey sandstone  
of Quebec Group. Locality: ME-7, St.  
Lawrence Group.

Photograph 84.

Sheet joints in impure grey sandstone of  
Quebec Group. Locality: ME-9, St.  
Lawrence coast.



Photograph  
83.



Photograph  
84.

Photograph 85.

Joints in Quebec Group Slate as seen on the dip slope of the  $S_2$  cleavage surface. They essentially consist of joints perpendicular to the  $S_s$  surface and trending parallel to the  $S_s$  dip. Locality: ME-1, St. Lawrence coast.

Photograph 86.

A dip slope of impure grey sandstone showing joint sets with varying relations to the dip of the rock. The dip, oblique and strike joints are respectively parallel, oblique and normal to the dip of the rocks. Locality: SW of ME-7, St. Lawrence coast.



Photograph  
85.



Photograph  
86.

Photograph 87.

Subhorizontal kink bands in the Quebec Group Slate. Locality: South of Portage Lake.

Photograph 88.

Subhorizontal kink bands in Fortin Group as seen on the ac joint surface. The displacement of the upper part of the kink is to the left, NW. Note smaller kink bands within the large one at  $15-20^{\circ}$  to the latter showing similar displacement. Locality: Caus 5/5 near Ste Florence village.



Photograph  
87.



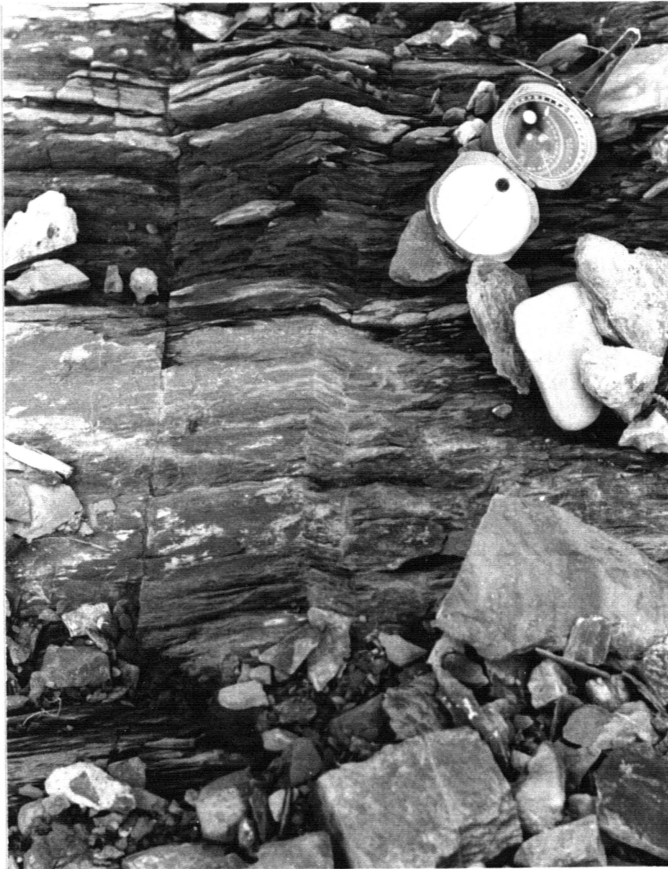
Photograph  
88.

Photograph 89.

Kink bands parallel to ac joints in Matapedia Group. Locality: Near St. Alexis, NW of Matapedia town.

Photograph 90.

Vertical view of small Z-shaped fold probably assignable to the  $F_3$  phase of folding.  
Locality: North of Mont Joli, St. Lawrence coast.



Photograph  
89.



Photograph  
90.

Photograph 91.

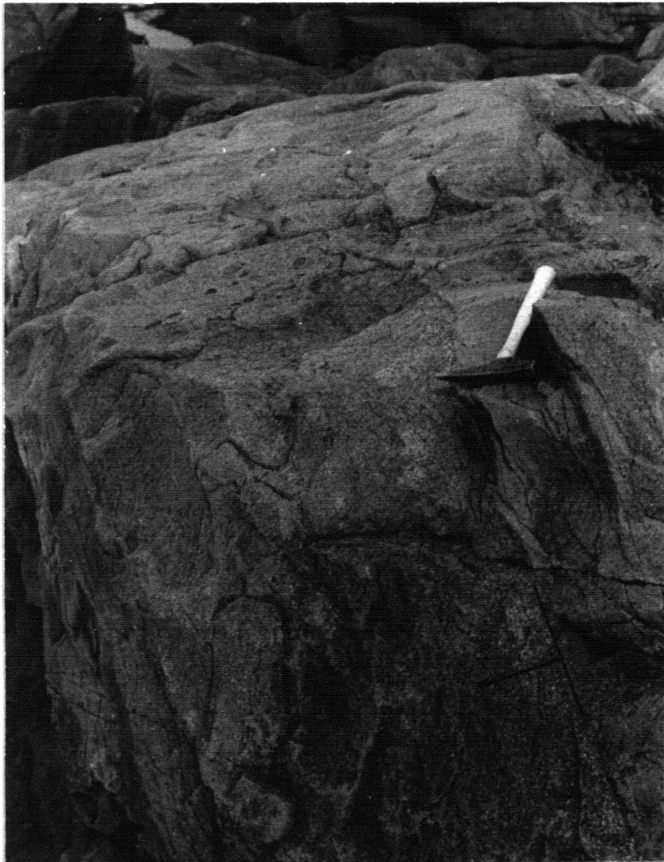
Intraformational conglomerate in Quebec Group Slate. The phenoclasts consists of locally derived, dense tabular limestone material. Locality: 3 miles of Riviere Matane village.

Photograph 92.

Inverted section of impure grey sandstone as shown by graded bedding. Locality: 5 miles NE of Ste. Felicite along St. Lawrence coast.



Photograph  
91.



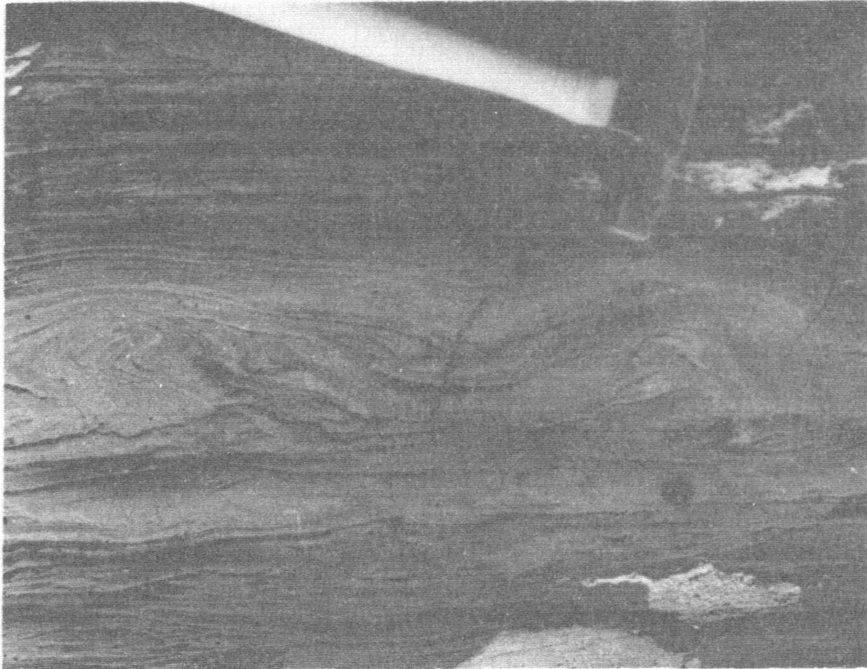
Photograph  
92.

Photograph 93.

Convolute bedding in calcareous siltstone layer in Quebec Group Slate. Locality: SW of Les Mechins.

Photograph 94.

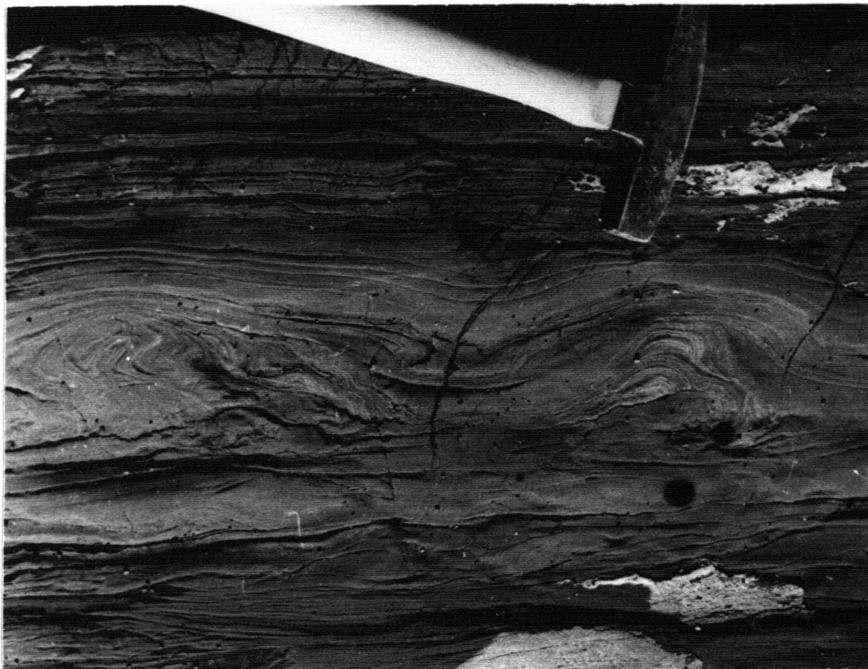
Calcite veins displaced by flexural slip between calcareous layers and minor shale layers in  $F_2$  concentric fold. Locality: ME-1, St. Lawrence coast.



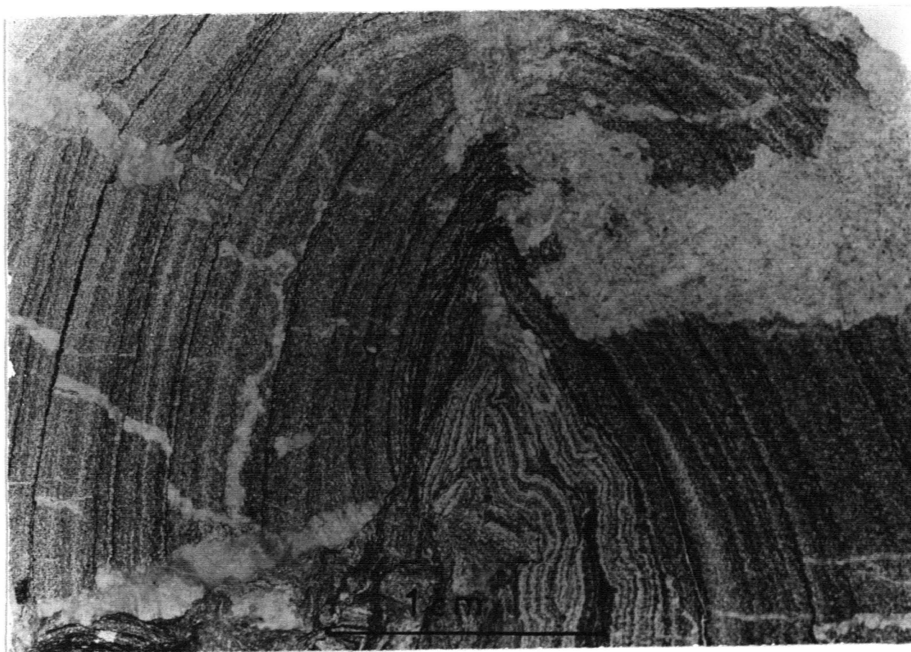
Photograph  
93.



Photograph  
94.



Photograph  
93.



Photograph  
94.

Photograph 95.

Disharmony in  $F_2$  folds in competent layers in a sequence of calcareous siltstone and shale of Quebec Group. Locality: ME-1, St. Lawrence coast.

Photograph 96.

Disharmony in  $F_2$  folded layers in Quebec Group Slate. Locality: St. Lawrence coast, ME 1.



Photograph  
95.



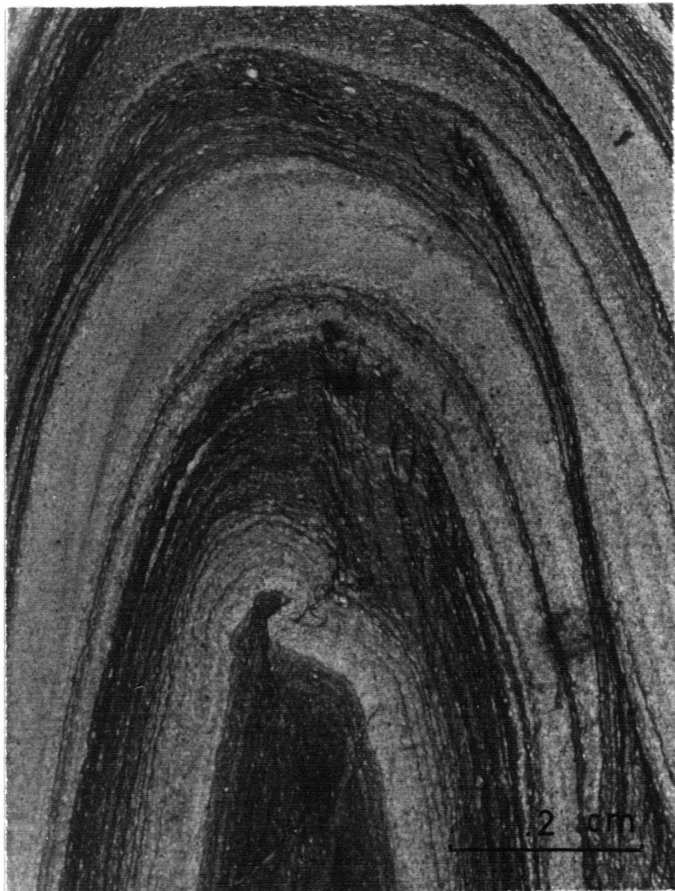
Photograph  
96.

Photograph 97.

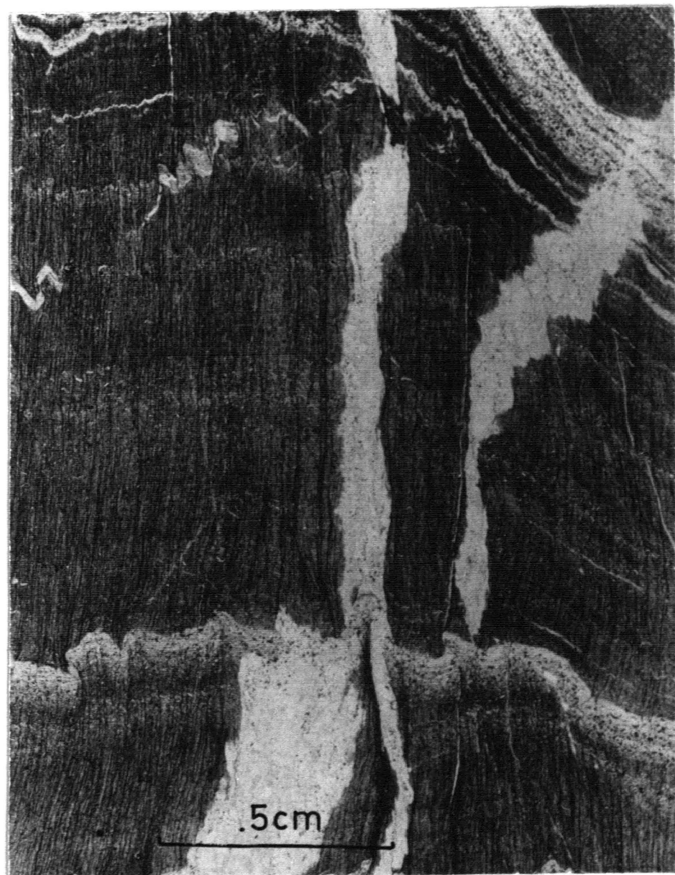
Isoclinal  $F_2$  fold in a shale and calcareous layers. Locality: St. Lawrence coast.

Photograph 98.

Calcite veins trending parallel to  $S_2$  cleavage.  
Locality: SW of Riviere Matane village.



Photograph  
97.



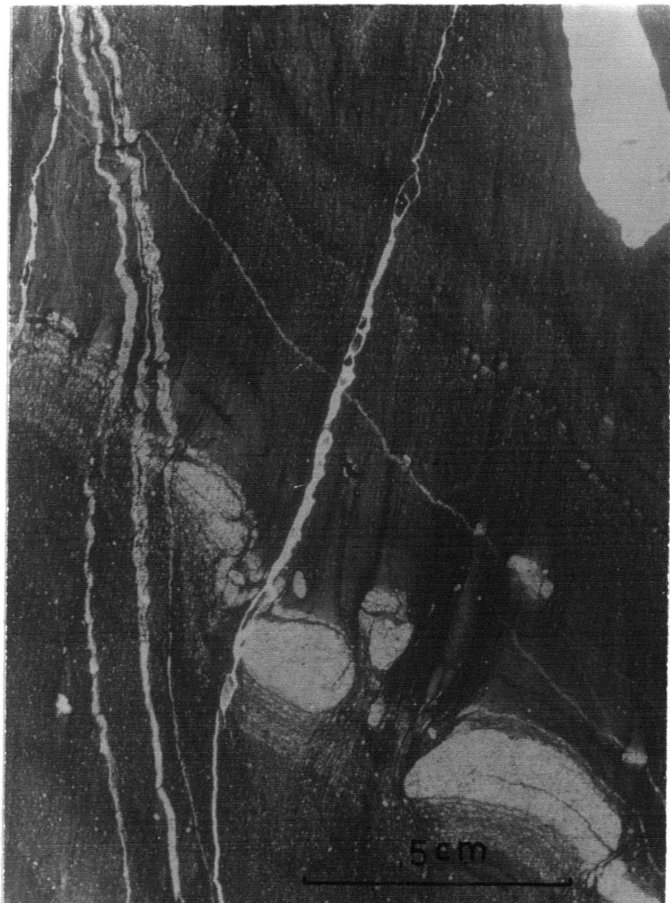
Photograph  
98.

Photograph 99.

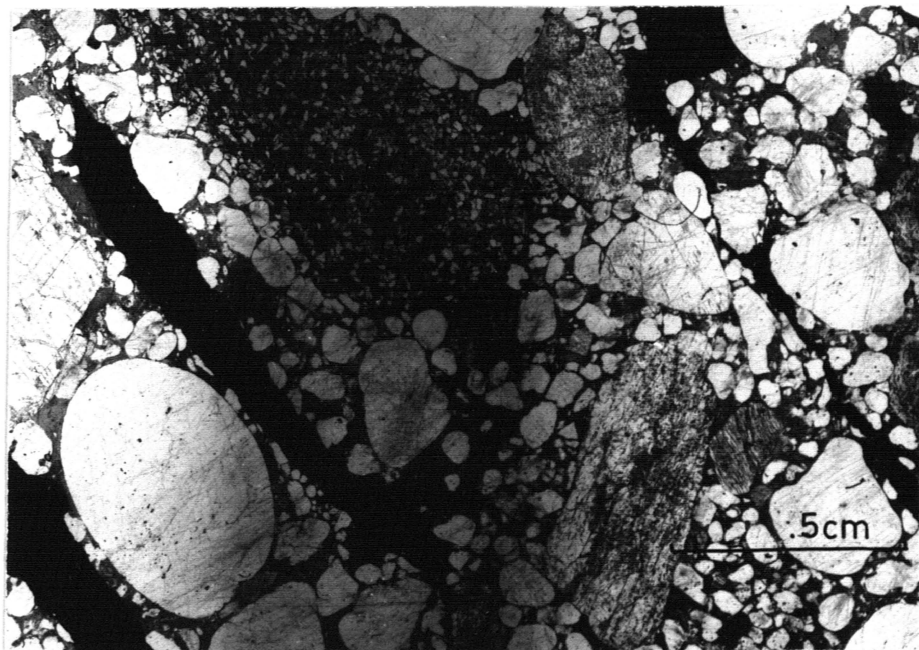
Discontinuous calcite layer separated into fragments by  $S_2$  in cleaved argillite in Quebec Group. The calcareous layer presumably was originally continuous. Locality: St. Lawrence coast north of Metis, MW-4.

Photograph 100.

Conglomerate in Ordovician Slate. Note well rounded quartz grains, limestone with relict bedding and limestone fragments with quartz peppered throughout. Elongated fragments of slate (black) lie parallel to the layering in the conglomerate. The enclosing slate is cleaved. The  $S_2$  cleavage is later than the deposition of the conglomerate. Locality: 6 miles west of Syabec,  $\frac{1}{2}$  mile south of St. Noel.



Photograph  
99.



Photograph  
100.

Microphotograph 1.

S<sub>1</sub> schistosity in the matrix of the meta-sedimentary rock of the Shickshock Group as defined by the preferred orientation of micaceous minerals. Locality: 4 miles east of Depot Jean, Shickshock Mountains.

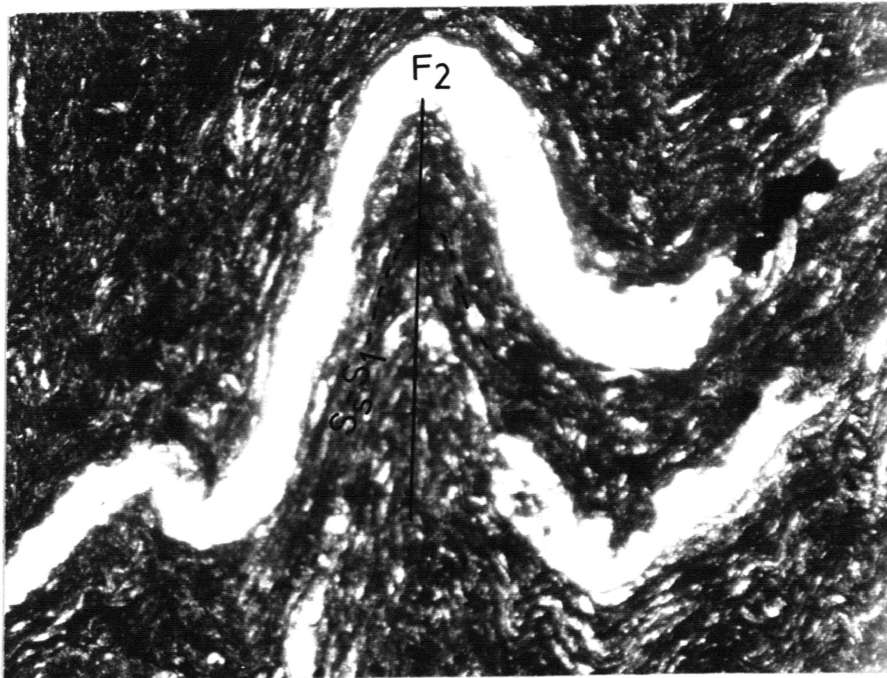
Microphotograph 2.

S<sub>1</sub> schistosity defined by preferred orientation of micas, chlorite and actinolite in metavolcanic rock of the Shickshock Group folded with quartz vein by F<sub>2</sub> folds.  
Locality: Shickshock Mountains.



Microphotograph  
1.

x80



Microphotograph  
2.

x80

Microphotograph 3.

Tight  $F_2$  fold deforming  $S_1$  schistosity and quartz vein in the Shickshock metavolcanic rock. Locality: Shickshock Mountains.

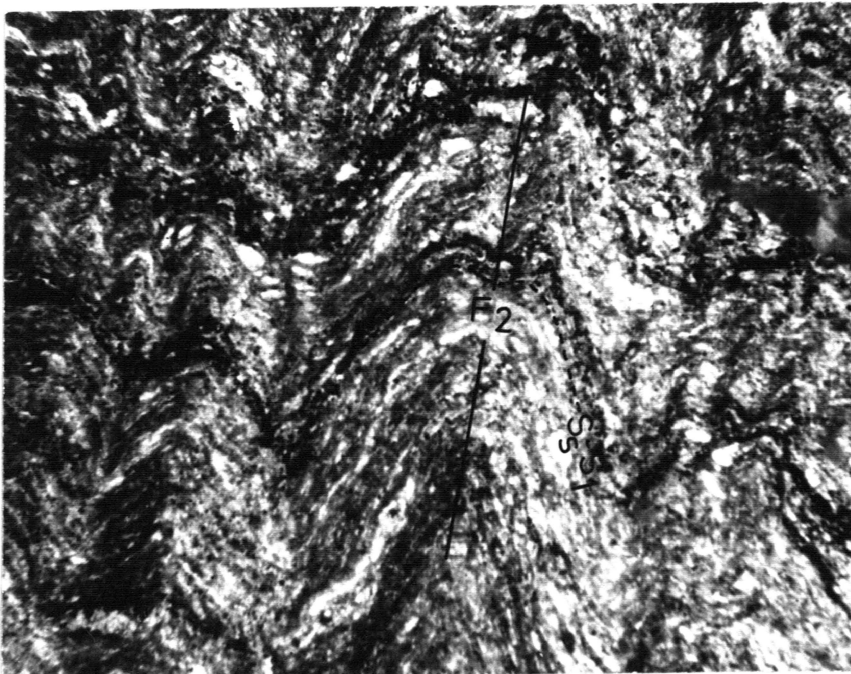
Microphotograph 4.

Symmetric  $F_2$  crenulation folds in Shickshock metavolcanic rock. Locality: Shickshock Mountains.



Microphotograph  
3.

x20



Microphotograph  
4.

x20

Microphotograph 5.

F<sub>2</sub> fold in calcareous siltstone and slate of Quebec Group. The middle layers are partly detached and appear punched upward. Locality: St. Lawrence coast.

Microphotograph 6.

F<sub>2</sub> fold in calcareous slate of Quebec Group demonstrates an increase in the frequency of S<sub>2</sub> axial plane cleavage on the limbs compared with the hinge area. Locality: 4 miles south of MW-3.



Microphotograph  
5.

x20



Microphotograph  
6.

x20

Microphotograph 7.

Bands with weakly developed  $S_2$  cleavage on the outer parts of hinges of successive  $F_2$  folds in calcareous layers within Slate of Quebec Group. Locality: ME-1, St. Lawrence coast.

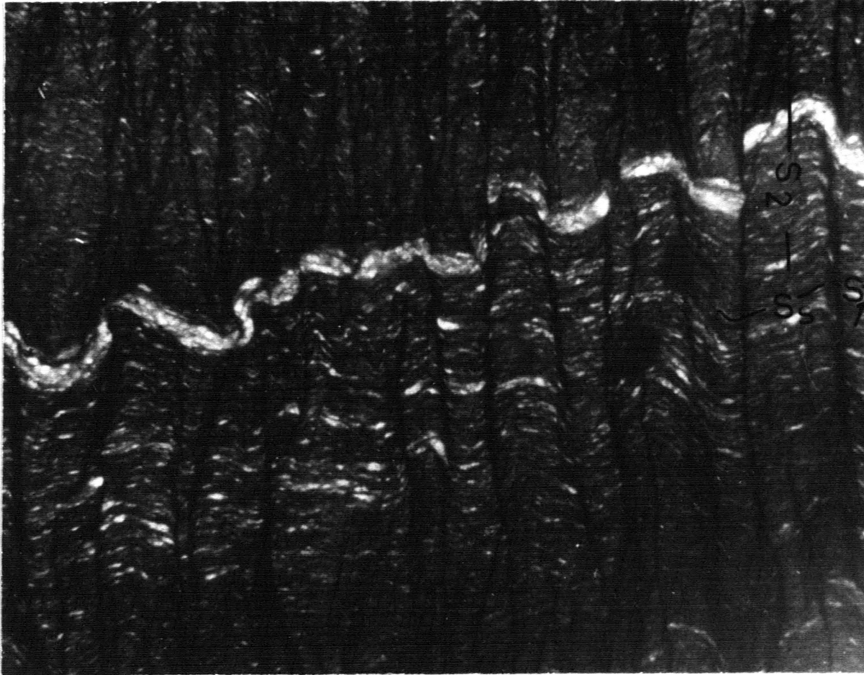


Microphotograph  
7.

x20

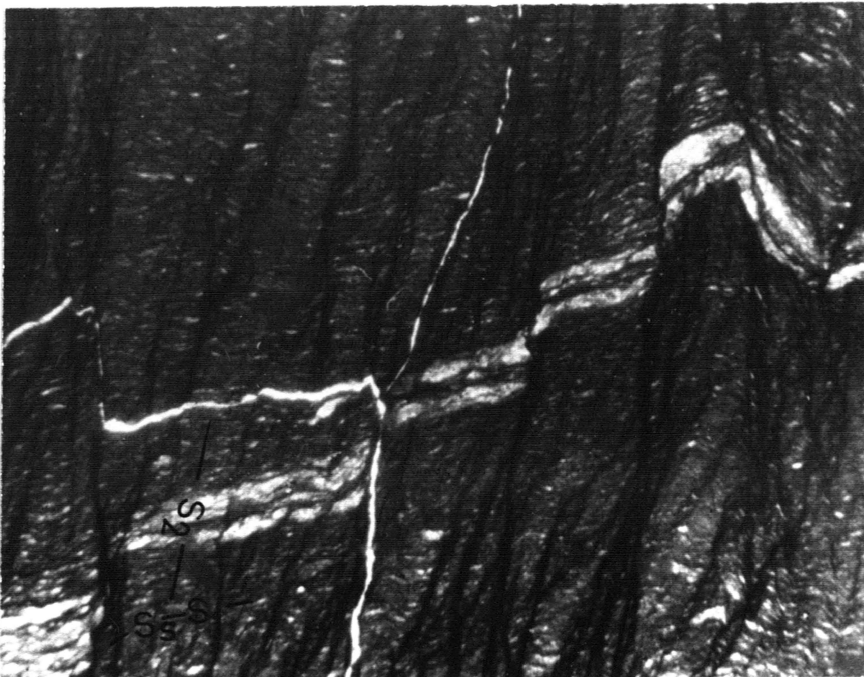
Microphotographs 8 & 9.

Offsets of thin calcareous layers across axial plane cleavage in Cambrian-Ordovician Slate. Locality: RM-1, 100 yards north of Riviere Matane village.



Microphotograph  
8.

x20



Microphotograph  
9.

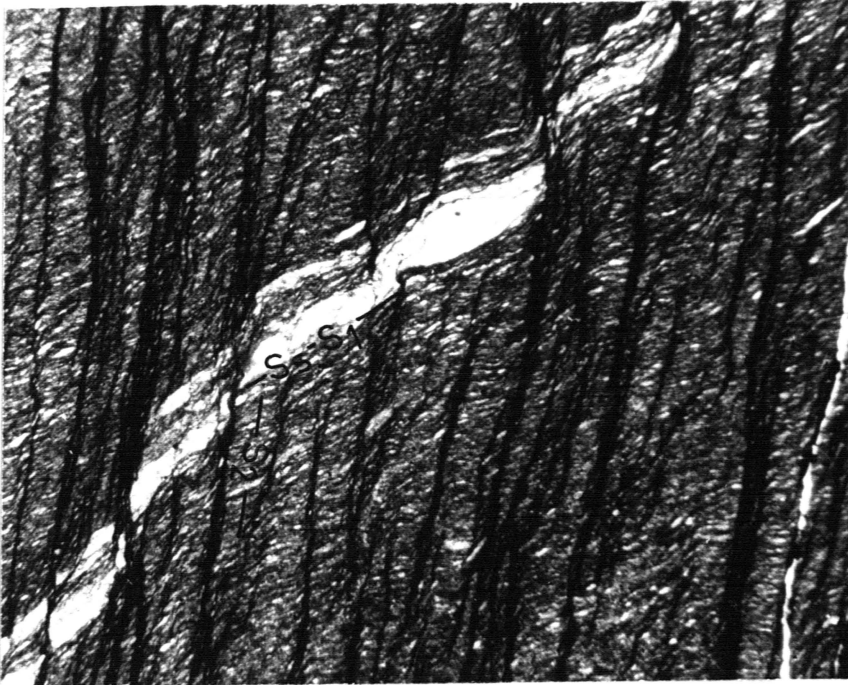
x20

Microphotographs 10 & 11.

Offsets across cleavage under varying magnification in Cambrian-Ordovician grey slates.

Locality: RM-1, 100 yards N of Riviere

Matane village.



Microphotograph  
10.

x20



Microphotograph  
11.

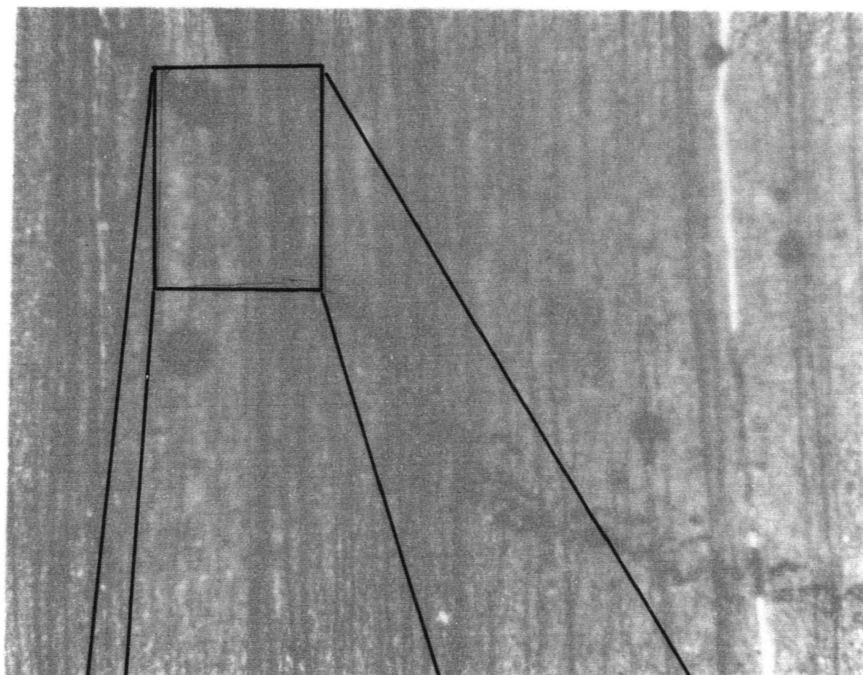
x80

Microphotograph 12.

S<sub>2</sub> cleavage as developed in the Quebec Group Slate. Locality: 4 miles south of MW-3.

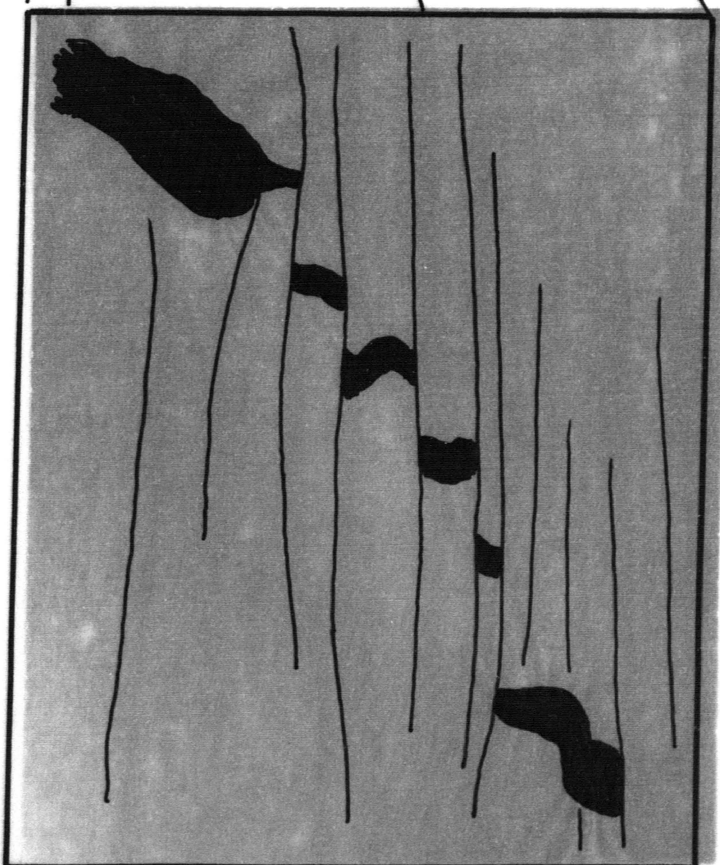
Microphotograph 13.

Offsets across S<sub>2</sub> cleavage planes under high magnification in acute F<sub>2</sub> fold in the Quebec Group Slate. Locality: 4 miles south of MW 3.



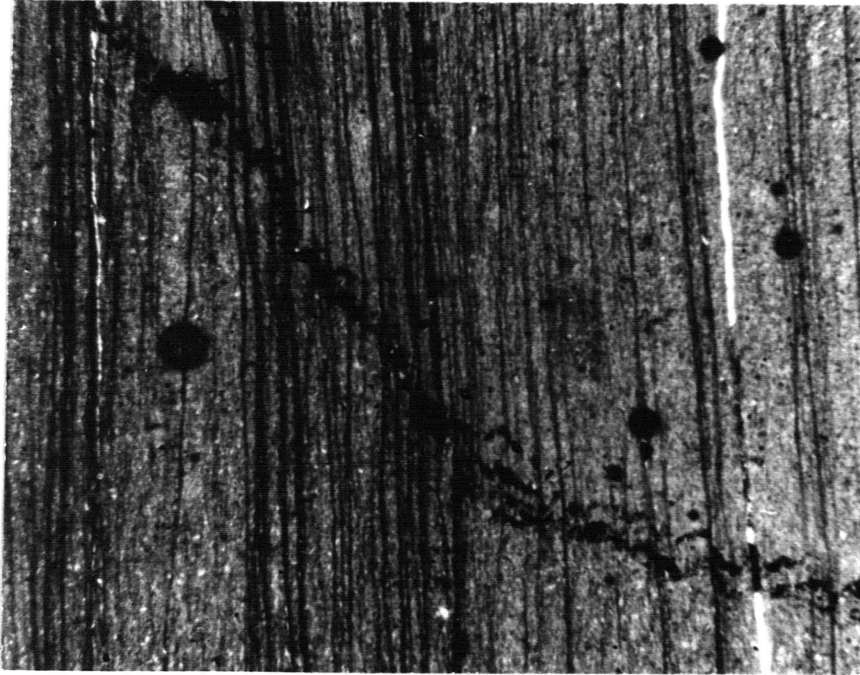
Microphotograph  
12.

x20



Microphotograph  
13.

x80



Microphotograph  
12.

x20



Microphotograph  
13.

x80

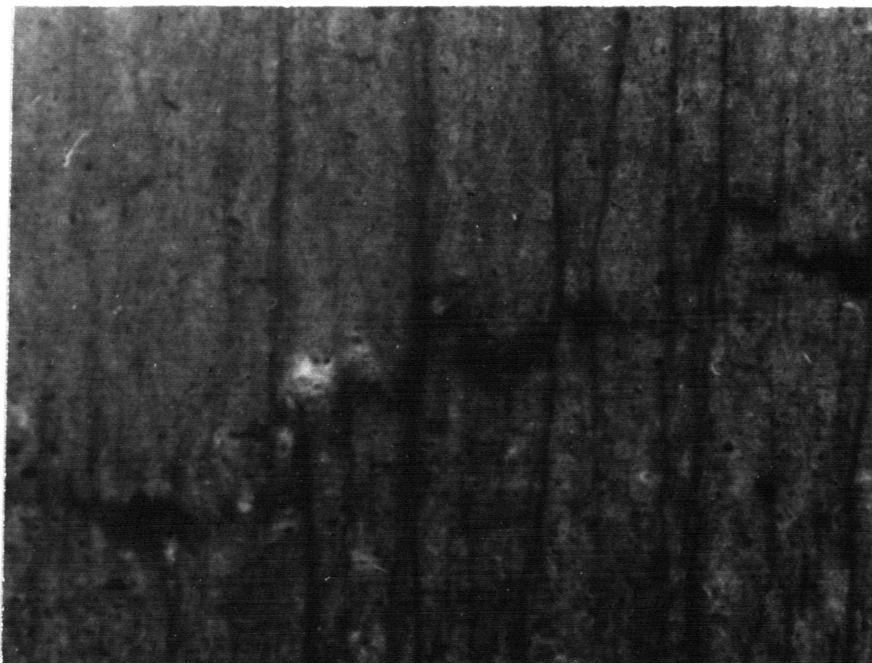
Microphotograph 14.

Offsets across  $S_2$  cleavage planes in an open  $F_2$  fold. The reference horizon is bedding marked by colour contrast, in Quebec Group Slate. Photographs 12, 13 and 14 demonstrate relationship between cleavage spacing and the angle between cleavage and bedding ( $S_2$   $S_S$ ).

Locality: 4 miles S of MW 3.

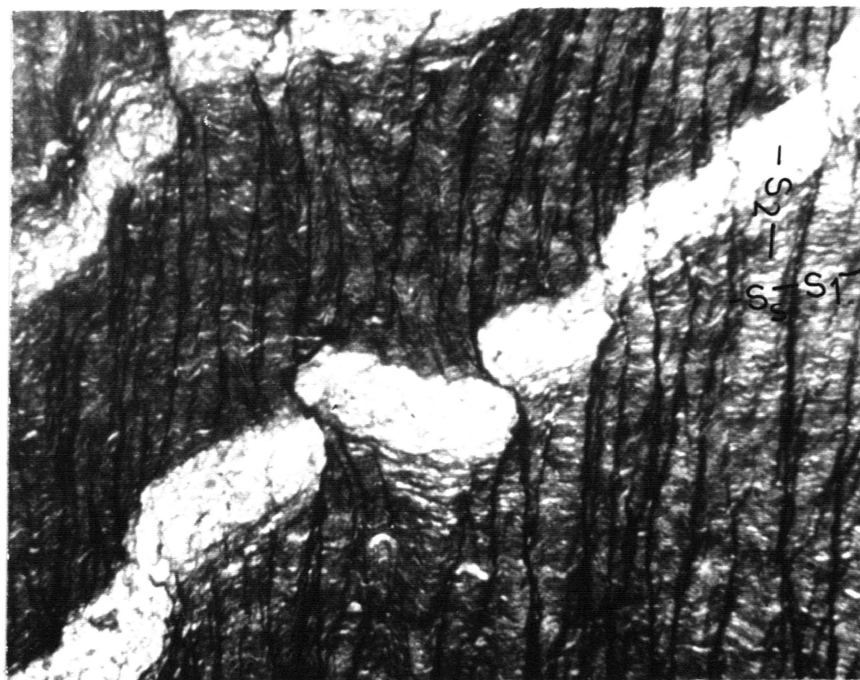
Microphotograph 15.

Deflection of  $S_2$  crenulation cleavage around fragment of a siltstone layer within Cambrian-Ordovician Slate. The deflection probably resulted from a greater shortening perpendicular to  $S_2$  of the slate than of the siltstone. A recrystallised mica fabric,  $S_1$  parallel to the layering is also present. Locality: RM-1.



Microphotograph  
14.

x80



Microphotograph  
15.

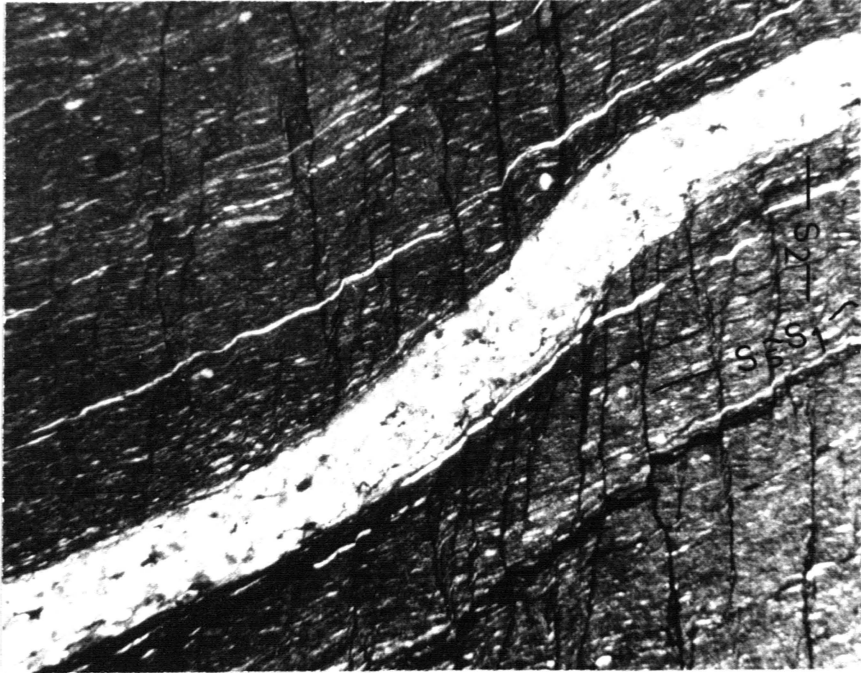
x20

Microphotograph 16.

Increase of  $S_2$  cleavage frequency with increased flexure in layering. The frequency of cleavage is inversely related to the angle between the layering and cleavage. The thick pale coloured layer is a calcite vein, slightly transverse to  $S_s$  and  $S_s$  in Cambrian-Ordovician grey slate. Locality: RM-1.

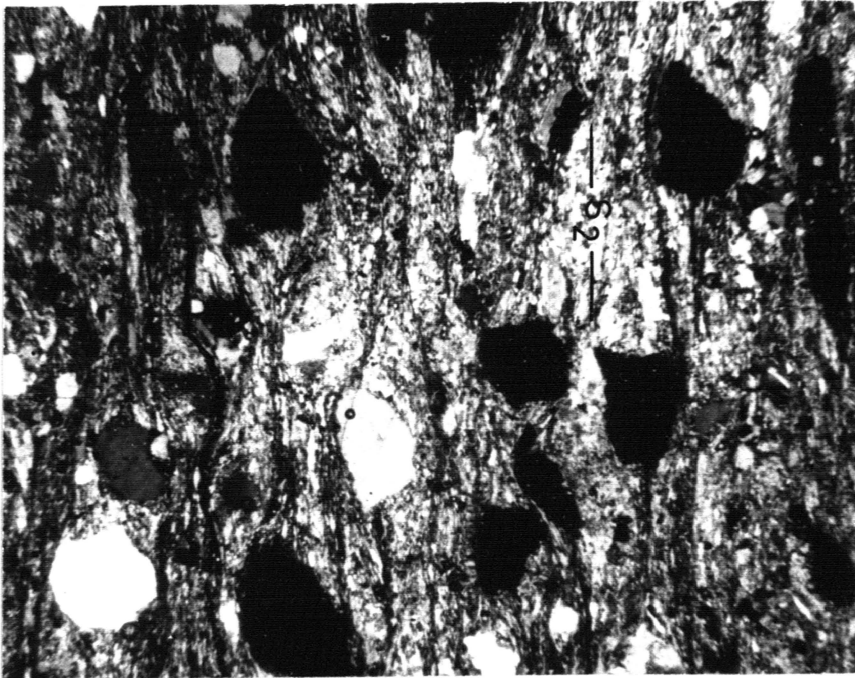
Microphotograph 17.

$S_2$  slaty cleavage in Fortin Group (Devonian). The cleavage bends around quartz grains. Locality: near Ste. Florence.



Microphotograph  
16.

x20



Microphotograph  
17.

x20

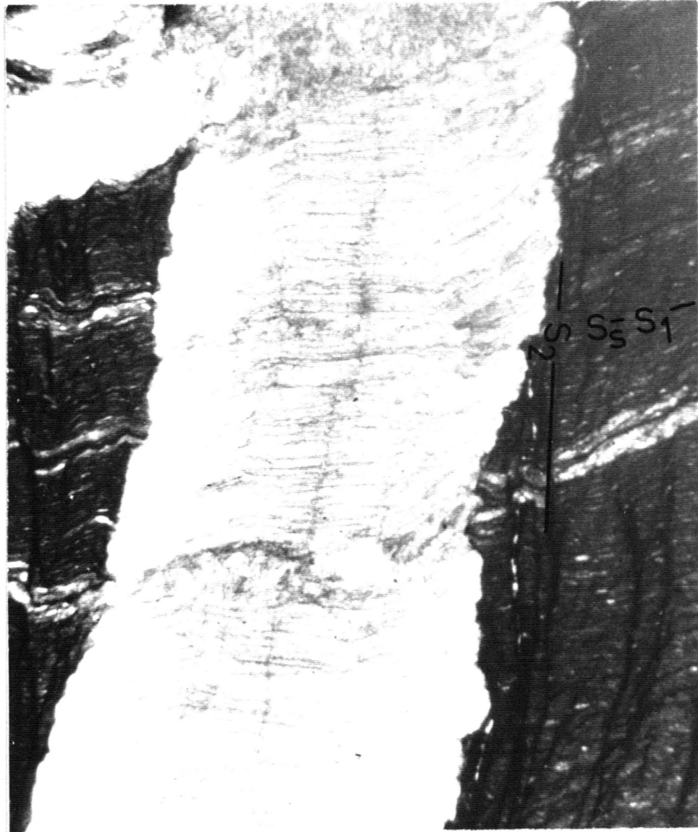
Microphotograph 18.

Calcite vein sub-parallel to the axial plane of the fold (in the sample, and not obvious in the microphotograph), discordant to the  $S_2$  fracture, in Cambrian-Ordovician grey slate.

Locality: RM-1.

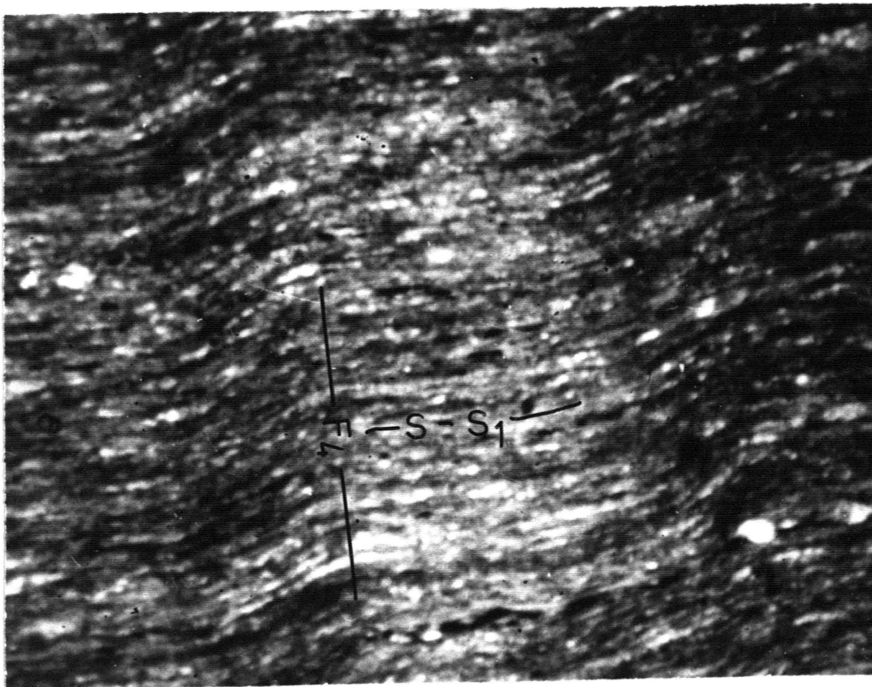
Microphotograph 19.

S-shaped  $F_4$  fold deforming the  $S_1$  schistosity in the Shickshock metavolcanics. Locality: Shickshock Mountains.



Microphotograph  
18.

x20



Microphotograph  
19.

x80

Microphotograph 20.

Impure grey sandstone of Quebec Group under polarised light. Sandstone contains basic volcanic rock fragments. Note poorly sorted chloritised detrital matrix, and absence of schistosity in the matrix. Locality ME-7, northeast of Ste. Felicite.

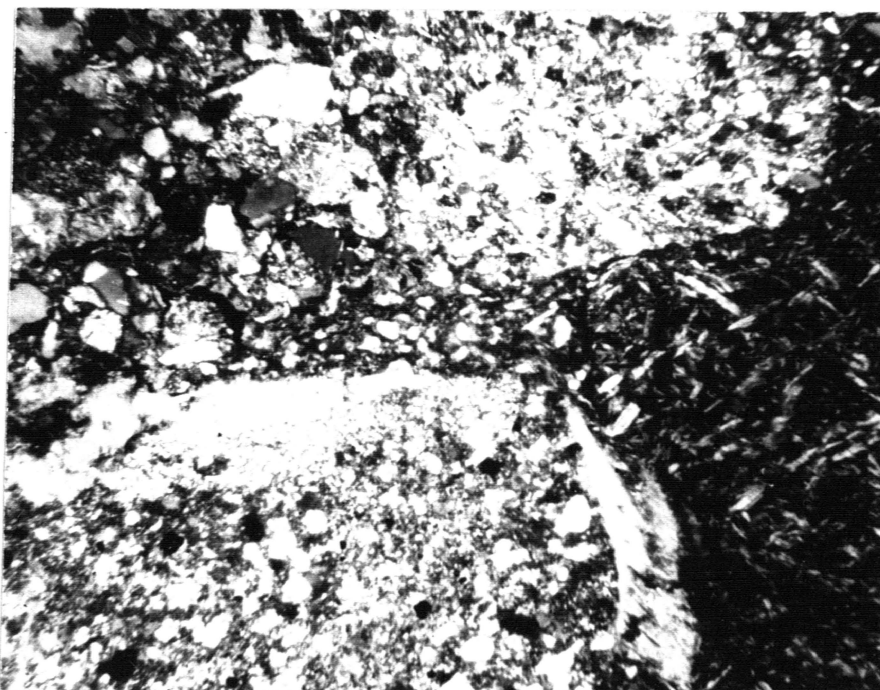
Microphotograph 21.

Same as above under cross nicols.



Microphotograph  
20.

x20

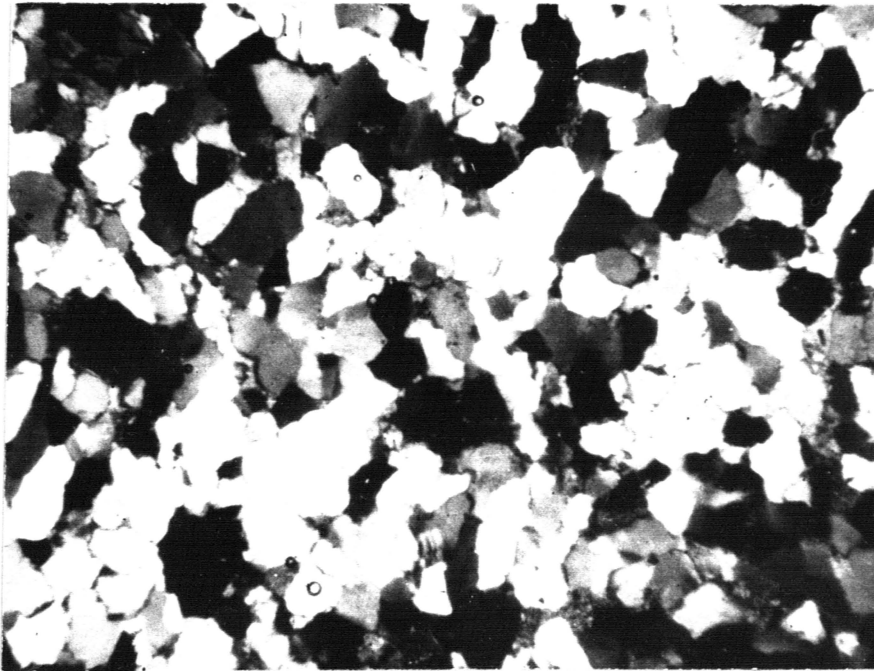


Microphotograph  
21.

x20

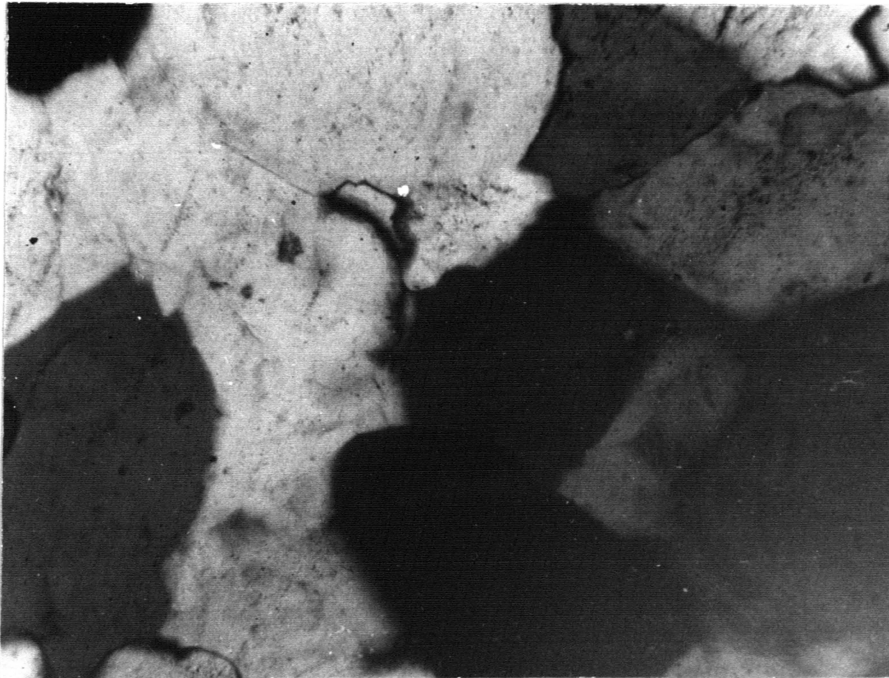
Microphotographs 22 and 23.

Kamouraska quartzite of Quebec Group at low and high magnification under cross nicols. Locality: 6 miles west of Riviere Matane village.



Microphotograph  
22.

x20



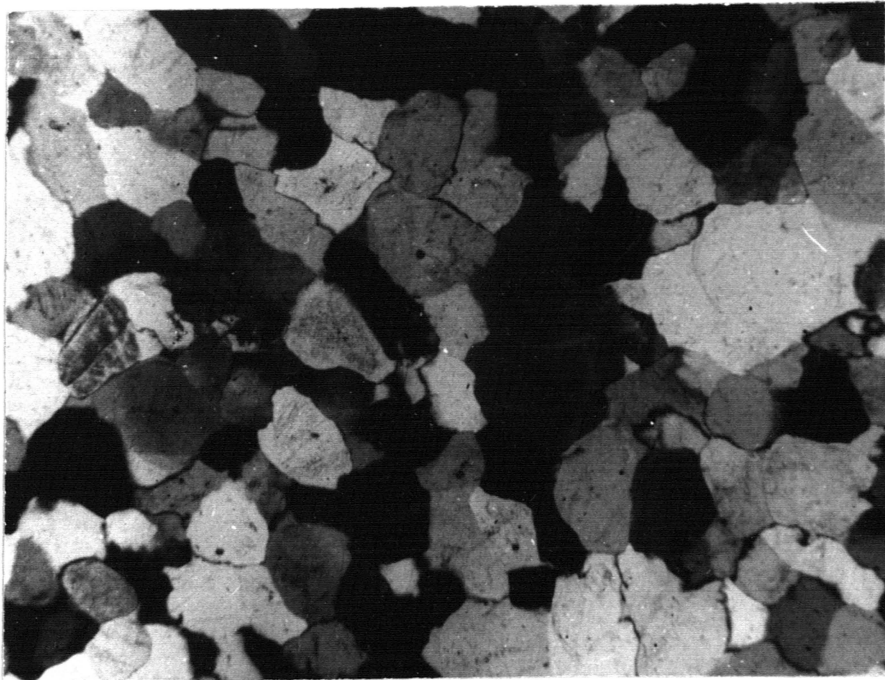
Microphotograph  
23.

x80

Microphotographs 24 and 25.

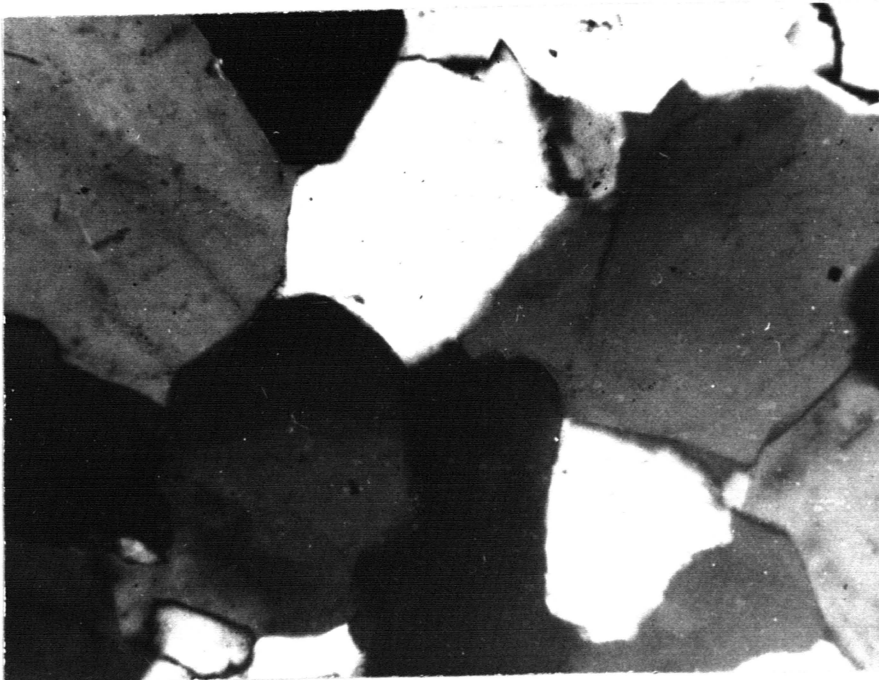
Val Brilliant quartzite of Middle Silurian age at  
low and high magnification under cross nicols.

Locality: 1 mile west of Val Brilliant village.



Microphotograph  
24.

x20



Microphotograph  
25.

x80