

STRUCTURAL GEOLOGY
OF LOWER PALEOZOIC ROCKS,
MT. ALBERT AREA GASPE PENINSULA,
QUEBEC

by
Alberto Carrara

A thesis
submitted in partial fulfillment of the requirements
for the degree of
DOCTOR OF PHILOSOPHY
in the
Department of Geology
University of Ottawa
April 1972

© Alberto Carrara, April 1972.

ABSTRACT

Lower Paleozoic volcanic and sedimentary rocks of the Shickshock and Quebec Groups in the area between Mt. Albert and the north shore of Gaspé Peninsula were folded by at least three phases of deformation during the Ordovician Taconic orogeny.

In the south the first phase of deformation produced a slaty to schistose foliation S_1 that lies parallel to the axial surfaces of isoclinal folds F_1 , and in most places parallel to bedding. Before later deformation the folds were probably recumbent and trending eastward. Emplacement of an ultramafic pluton at Mt. Albert and metamorphism up to the upper greenschist facies accompanied the deformation. This is considered to be an early Middle Ordovician event because F_1 folds affect sedimentary rocks of this age, and biotite and hornblende aligned along S_1 foliation in the Shickshock metavolcanics, yielded concordant K-Ar ages (averaging 465 m.y.).

The F_1 structures, such as S_1 , are absent to the north where the rocks are unmetamorphosed. The boundary between slates with S_1 parallel to bedding and adjacent shaly rocks with a morphologically similar sedimentary parting is obscure; identification was aided by determining the preferred orientations of chlorite by X-ray diffraction. The boundary, which is not marked by a recognizable unconformity, is defined as the upper limit of F_1 structures confined to lower levels of the stratigraphic pile.

The second deformation produced inclined to upright, northeast trending F_2 folds, which are present in all the Lower Paleozoic rocks of the area. Little or no metamorphism was associated with this deformation.

Axial plane cleavages of F_2 folds include slaty, fracture, crenulation and intermediate types. The character is dependent upon the lithology,

amount of strain, presence or absence of a previous fabric such as S_1 , and the amount of mineral differentiation occurring during the formation of cleavage. These types of cleavage also occur with folds younger than F_2 , hence they are not typical for a particular generation. However, a truly penetrative slaty cleavage is unique to S_1 foliation.

Changes in thickness of layers around F_2 folds suggest that the dominant folding mechanism was buckling accompanied by a component of homogeneous strain (flattening). In most competent quartz or calcite-rich layers interbedded in pelites, flattening was limited (not more than 20%), but widespread. No correlation was observed between flattening and interlimb angle, wavelength, or layer thickness. As in buckle folds measured elsewhere, the wavelength is almost directly a function of the dominant member.

The third phase of deformation produced only local structures. These include upright folds with variably spaced fracture cleavage near the shore of the St. Lawrence River and steeply dipping kink bands in the rocks of the interior foliated by S_1 . The folds and kinks both trend northwest.

During the mid-Devonian Acadian orogeny, the Lower Paleozoic rocks and Silurian-Devonian formations, now in faulted contact to the south of Mt. Albert, were gently folded along east-west trending axes, a trend that dominates the present outcrop pattern. A late Acadian granitic pluton was emplaced into Lower Paleozoic rocks to the south.

Preliminary investigation of Lower Paleozoic rocks elsewhere in northwestern Gaspé indicates that in a large part of the northern belt of the peninsula the succession of structures is similar to that established in the Mt. Albert area.

	TABLE OF CONTENTS	PAGE
I	INTRODUCTION	
	Location, Accessibility and Type of Outcrop.....	1
	Geological Setting.....	3
	Previous Work.....	4
	Problems.....	9
	Purpose of the Present Study.....	10
	Acknowledgements.....	11
II	STRATIGRAPHY	
	Introduction.....	12
	Cambro-Ordovician Volcanic and Sedimentary Rocks.....	12
	Shickshock Group.....	12
	Quebec Group.....	17
	Phyllitic-slate Unit.....	17
	Argillite-greywacke Unit.....	19
	Siltstone Unit.....	20
	Silurian Sedimentary Rocks.....	21
	Devonian Sedimentary and Volcanic Rocks.....	22
	Igneous Rocks.....	22
	Mt. Albert Serpentinite.....	22
	Table Top Granite.....	23
III	STRUCTURAL PHASES	
	Introduction.....	25

D ₁ Deformation.....	28
F ₁ Folds.....	28
L ₁ Lineations.....	32
S ₁ Foliations.....	33
Boundary of F ₁ Structures.....	34
Age of Deformation.....	35
D ₂ Deformation.....	36
F _{2a} Folds.....	37
F _{2b} Folds.....	37
S ₂ Cleavage.....	40
L ₂ Lineations.....	41
Orientation.....	42
Strain in Deformed Sandstone Dykes.....	43
Age of Deformation.....	45
D ₃ Deformation.....	47
F ₃ Folds.....	47
Orientation.....	47
Age of Deformation.....	49
D ₄ Deformation.....	50
F ₄ Folds.....	50
Age of Deformation.....	50
General Structural Interpretation.....	51
Faults.....	52
Summary and Problems.....	53

IV METAMORPHISM AND MAGMATISM

Introduction.....	56
-------------------	----

Regional Metamorphism.....	56
Metamorphism and Deformation.....	58
Radiometric Ages.....	60
Contact Aureole of Mt. Albert Serpentinite.....	63
Age of Intrusion.....	64
Contact Aureole of Table Top Granite.....	69

V FOLD ANALYSIS

Introduction.....	70
Procedure.....	71
Field Classification.....	75
Laboratory Measurements.....	77
Fold Classification and Lithology.....	77
General Fold Mechanism.....	79
Wavelength and Layer Thickness.....	80
Viscosity Ratio and Wavelength-Layer Thickness Ratio....	85
Flattening.....	86
Interlimb Angle.....	90
Relationships between Flattening, Interlimb Angle and Other Parameters.....	90
Comparison of Interlimb Angles and Flattening between F_1 and F_2 Folds.....	93
Areal Variation of Interlimb Angle and Flattening in F_2 Folds.....	95
Summary.....	99

VI CLEAVAGE TYPES IN SEDIMENTARY AND LOW GRADE METAMORPHIC ROCKS

Introduction.....	101
-------------------	-----

	Slaty Cleavage Group.....	103
	Fracture Cleavage Group.....	106
	Crenulation Cleavage Group.....	106
	Cleavage Distribution.....	108
	Cleavage Type and Fold Generation.....	110
	Discussion.....	110
VII	ANALYSIS OF PREFERRED ORIENTATION OF CHLORITE BY MEANS OF X-RAY DIFFRACTION	
	Introduction.....	114
	Previous Work.....	114
	Present Work.....	116
	Experimental Procedure.....	117
	Limitations of Method and Sources of Error.....	119
	Data Treatment.....	121
	Discussion.....	127
	Preferred Orientation and Strain.....	130
	Summary.....	132
VIII	SYNTHESIS	
	Conclusions on Structures and Fabrics.....	134
	Tectonic History.....	135
	Structural History of Lower Paleozoic Rocks, Gaspé Peninsula	138
	Tectonic Speculations.....	144
IX	SUGGESTIONS FOR FUTURE WORK.....	146

REFERENCES.....	148
APPENDIX I K-AR AGES FROM MT. ALBERT AREA.....	162
APPENDIX II EQUAL AREA PROJECTIONS.....	163
Explanations of the Stereonets.....	164
APPENDIX III FOLD ANALYSIS	
Linear Regression.....	183
Correlation Coefficient.....	184
APPENDIX IV PRELIMINARY STRUCTURAL INVESTIGATION IN TEMISCOUATA AREA	
Introduction.....	189
Stratigraphy.....	190
Quebec Group.....	190
Silurian Formations.....	190
Devonian Formations.....	191
Structural Geology.....	193
Conclusions.....	195
APPENDIX V PLATES.....	197

	LIST OF FIGURES	PAGE
1	a) Location of the study area and of previous structural investigations in Gaspé Peninsula.....	2
	b) Previous geological maps that include the Mt. Albert area.....	2
2	Lithological map of the Mt. Albert area.....	15
3	Schematic cross-section of the Mt. Albert area.....	16
4	Geological map and interpretative cross-section of the Mt. Albert area.....	in pocket
5	Generalized structural map of the Mt. Albert area.....	27
6	F_1 fold profiles.....	29
7	Epidote knot with adjacent F_1 and F_2 folds.....	31
8	Profiles of small-scale F_2 folds.....	39
9	Strain in deformed sandstone dykes.....	44
10	F_3 folds.....	48
11	Time-relations between deformation and metamorphism.....	61
12	Garnet crystals in metasediments within Mt. Albert contact aureole.....	65
13	Features of crystal growth during contact metamorphism and metamorphic differentiation accompanying deformation.	67
14	Classification and graphical representation of fold shapes in profile (Ramsay, 1967).....	72
15	Graph showing relationships between layer thickness variations and flattening strain in fold profiles.....	74
16	Relative frequency of Ramsay's fold classes observed in folded layers of the study area.....	78
17	Relative frequency of ratios of quarterwavelength, to layer thickness.....	82
18	Logarithmic relationship of fold quarterwavelength to thickness of dominant members.....	84

19	Relative frequencies of interlimb angles and percentage of flattening in folded layers in F_1 and F_2 fold specimens.....	87
20	F_2 fold profile and graph of t' versus dip α for calcite and pelite-rich layers.....	89
21	Relationship between percentage of flattening and interlimb angle.....	91
22	Relationship between percentage of flattening and layer thickness.....	92
23	Percentage of flattening and interlimb angles in F_1 and F_2 folds.....	94
24	Map showing the areal variation in interlimb angle of small-scale F_2 folds.....	96
25	Map showing the areal variation in percentage of flattening of small-scale F_2 folds.....	97
26	Cleavage types in sedimentary and low-grade semi-pelites.....	104
27	Cleavage types in sedimentary and low-grade pelites and semi-pelites.....	107
28	Schematic representation of the cleavage types in the sedimentary and low-grade pelites and semi-pelites of the study area.....	109
29	Map of Mt. Albert area with preferred orientation indices of chlorite.....	124
30	Histograms of preferred orientation index for chlorite...	126
31	Sketch of partly exposed F_1 fold in slates and siltstones of the Quebec Group.....	129
32	Areal distribution of the Taconic and Acadian major structures in Gaspé Peninsula.....	139
33	Stereo-net projection of L_1 lineations in the Cambro-Ordovician rocks of domain 1.....	166
34	Stereo-net projection of S_2 and L_2 in all Cambro-Ordovician rocks of the Mt. Albert area.....	167-8
35	Map showing S_2 and L_2 subareas.....	169
36	Stereo-net projections of S_2 and L_2 in subareas I-V.....	170-9

37	Map showing S_3 cleavage subareas.....	180
38	Stereo-net projections of S_3 in subareas I and II.....	181-2
39	Geological map of Témiscouta area.....	192

LIST OF TABLES

1	Rock Units.....	13
2	Characters of Fold Generations.....	26
3	Preferred Orientation Index (POI) for Chlorite.....	122
4	Comparison of Structural Successions in Three Areas of Cambro-Ordovician Rocks, North Gaspé.....	141
5	Lithology, Fold Class, Interlimb Angle, and Flattening of Folded Layers from the Study Area.....	185-7
6	Average Interlimb Angle and Flattening for F_1 and F_2 Folds from the Study Area.....	188

LIST OF PLATES

1	Structures and Textures in the Shickshock Metavolcanics..	197
2	D_1 and D_2 Structural Features.....	198
3	F_2 folds in Coastal Exposures of Argillite-Greywacke Unit.....	199
4	Small-scale F_2 Disharmonic Folding.....	200
5	D_3 Deformation and D_2 Segregation Features.....	201
6	Garnet Crystals in Metasediments within Mt. Albert Contact Aureole.....	202
7	Cleavage Types in Sedimentary and Low-Grade Pelites and Semi-pelites.....	203
8	Cleavage Types in Low-Grade Pelites.....	204

9	Cleavage Types in Sedimentary and Low-Grade Pelites and Semi-pelites.....	205
10	Cleavages in Siluro-Devonian Formations of Témiscouata areas.....	206

INTRODUCTION

Gaspé Peninsula in the northern part of the Appalachian mountain system is underlain by Paleozoic rocks that form an elongated, north-east to east trending, synclinorium. Devonian and Silurian formations are exposed along the central part of the peninsula, flanked to the north by Cambro-Ordovician and to the south by Ordovician-Silurian rocks. The major structure, known as the Gaspé-Connecticut Valley synclinorium, extends as far as southern New England, a distance of more than 1000 km.

Location, Accessibility and Type of Outcrop

The study area is in North Gaspé, on the northern margin of the Gaspé synclinorium. It extends from the shore of St. Lawrence River, east of Ste-Ann-des-Monts, to the south of Mt. Albert (Fig. 1a,b). The area is bounded by longitudes $66^{\circ}30'$ and $66^{\circ}00'$, and latitudes $48^{\circ}49'$ and $49^{\circ}14'$; it is 30-45 km from north to south and 35 km from east to west and covers approximately 1200 square kilometers.

Most of the area is in Tourelle, Christie, Courcellette, and La Potardière townships, covered by the Ste-Anne-des-Monts and Mt. Albert topographic sheets (22G/1 and 22B/16) of the Canadian National Topographic 1:50,000 Series. For convenience, after the well-known mountain in the south, it is here referred to as the Mt. Albert area.

Easy access to the interior is provided by the Trans-Gaspesian highway which crosses the area from northwest to southeast. Exposures along this highway were found particularly useful because thick vegetation covers most of the area. Moreover, large parts of the valley sides are covered by glacial till.

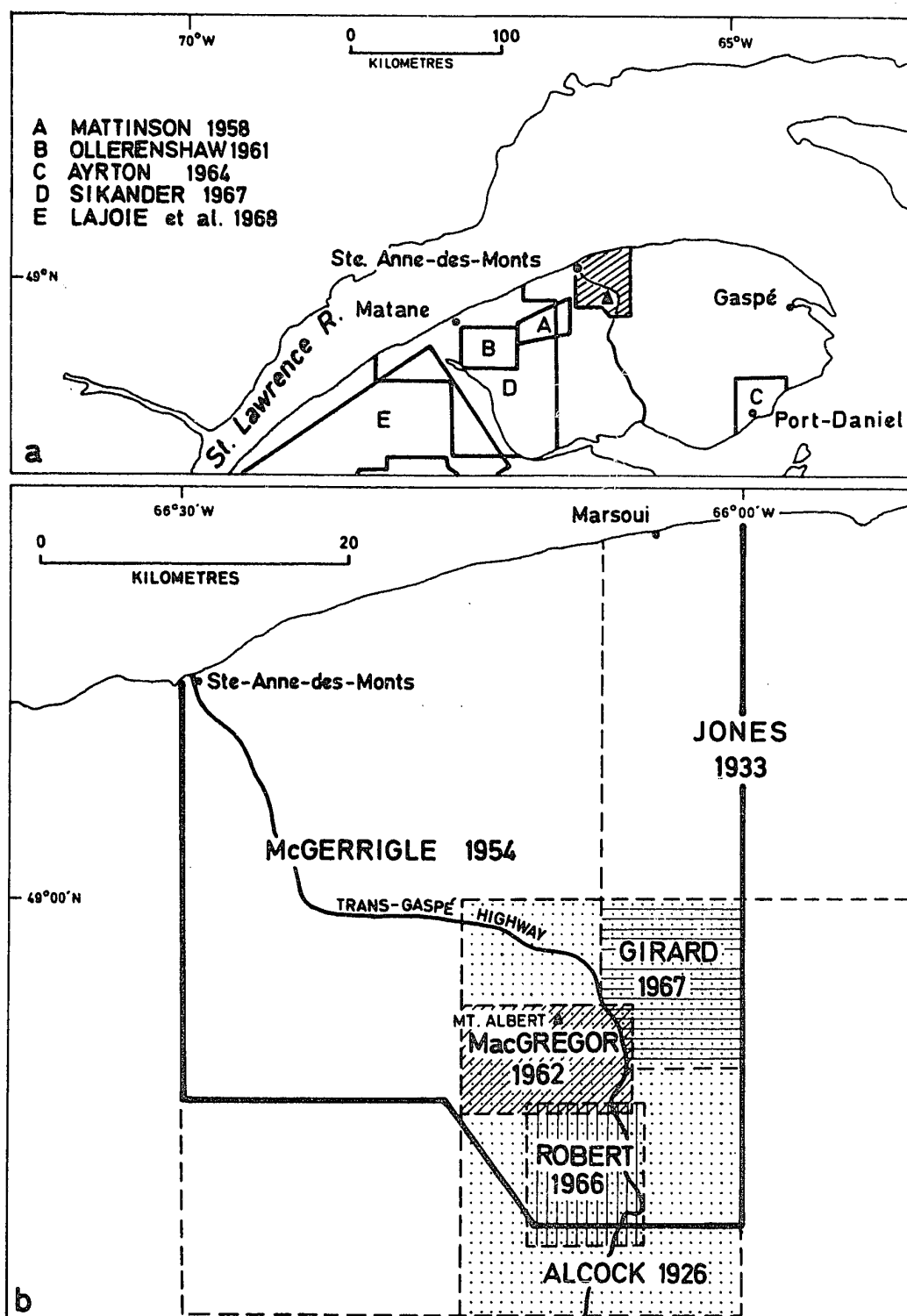


Fig. 1 a) Location of the study area and of previous structural investigations in Gaspé Peninsula.
b) Previous geological maps that include the Mt. Albert area.

Along gravel roads that cross the area from north to south, and a network of lumber roads, outcrops are generally small and not suitable for structural investigation. The best natural outcrops inland are along streams and rivers; elsewhere bedrock exposures in the interior are poor and occupy less than 3% of the total surface. In contrast, the shore of the St. Lawrence offers about 40 km of almost continuous outcrop.

The area varies in elevation from sea level at the St. Lawrence shore to 1000 m in the Shickshock range in the south (Mt. Albert rises to 1250 m). Intervening between the coast and the Shickshocks is an upland with elevations up to 950 m, which is crossed to the north of the Shickshocks by the east-west valley of Ste - Anne river.

Two field seasons (1968-1969) were spent in the area, and three weeks in the summer of 1970 were used to check previous observations. In 1971 five weeks were spent in the Témiscouata area, 200 km southwest of Mt. Albert, during a preliminary structural investigation.

All locations and sample numbers cited in the figures, tables, and plates of this work, are listed on a map at 1:50,000 deposited at the Department of Geology of the University of Ottawa.

Geological Setting

Gaspé synclinorium can be divided into two structural units: a narrow northern belt, approximately 40 km wide and 500 km long, which extends from Rivière-du-Loup to the eastern tip of the peninsula, and a much broader southern belt embracing most of the rest of Gaspé Peninsula (Fig. 1a, 32).

The northern belt is characterized by a thick, severely deformed sequence of flysch or flysch-like Cambro-Ordovician shales, greywacke, conglomerate, limestone (Quebec Group), associated on the southern margin with metavolcanics (Shickshock Group) and ultramafic intrusions as at Mt. Albert.

A thick sequence of platform-type sandstones, conglomerate, and limestone of Siluro-Devonian age predominates in the southern belt. These rocks are separated from the older northern rocks by a high angle fault (Shickshock fault) or by an unconformity. Rocks of the southern belt were gently folded along the central part of the peninsula and more tightly folded southward, during the mid-Devonian Acadian orogeny.

In the south, Silurian sedimentary rocks (Matapedia Group) conformably overlie Ordovician deposits (Pavlides et al., 1968) and only in the Port-Daniel area (Fig. 1a, 32), is there evidence of an angular unconformity between Ordovician and Silurian rocks.

Previous Work

The first geological observations in Gaspé date back to the last century when Logan (1846), Murray (1947), and Richardson (1859) examined the rocks along the coast and inland. The sedimentary rocks of the north shore were correlated with those that outcrop near Quebec City (Quebec Group) and a structural section across the peninsula was drawn by Logan (1863).

Ells (1885) and Low (1885) continued the survey of the center of the peninsula and published the first observations of the Mt. Albert serpentinite intrusion.

Forty years later, Alcock (1924) studied part of the Matane area and Shickshock Mountains. The Mt. Albert intrusive, the adjacent metavolcanic and Silurian sedimentary rocks were also mapped by Alcock (1926), and Jones (1933) mapped the Table Top granite body and the Marsoui area immediately to the east of the study area (Fig. 1b).

A regional synthesis of central western Gaspé was carried out by Crickmay (1932), but the most comprehensive contributions to the understanding of the stratigraphy of Gaspé were by McGerrigle (1950, 1953, 1954), and the regional map based on his and other material (McGerrigle and Skidmore, 1967) forms an essential background to later work.

As regards the study area, McGerrigle (1954) surveyed the Tourelle and Courcellette areas, located northwest and west of Mt. Albert, (Fig. 1b) and gave a detailed description of the different formations outcropping in the region. On the basis of paleontological evidence, confirmed with few exceptions by Riva (1968, written communication, 1968), the northern part of the Quebec Group was divided into the Deepkill (Lower Ordovician) and Normanskill (Middle Ordovician) Formations. The part of this Group adjacent to the Shickshock rocks did not bear fossils and was called "Lower Ordovician or Older". McGerrigle suggested that the Shickshock metavolcanics, because of their complex structure and metamorphism, were probably Precambrian.

Tanner and Uffen (1960) delineated an axis of abnormally high gravity that extends from the Shickshock mountains east to Gaspé Bay. If this anomaly is due to the presence of the basic volcanic and ultramafic rocks of the Shickshock Group, then these rocks extend eastward beneath the younger Silurian-Devonian cover. Following McGerrigle (1954), these authors referred to the Shickshock rocks as Precambrian.

The first detailed petrological study was by Mattinson (1958, 1964), who noted in the Mt. Logan area (Fig. 1a) a gradual increase in metamorphic grade from unmetamorphosed rocks along the shore of the St. Lawrence to greenschists with garnet on the southern edge of the Shickshock mountains. Thus the metamorphism of the Shickshock rocks cannot be ascribed to Precambrian events.

Ollerenshaw (1961, 1963, 1967) mapped the western part of the Shickshock Mountains in the Couq-Longis area (50 km west of the study area, Fig. 1a), and considered that the Shickshock metavolcanics form a wedge, with interfingering contact, within the Quebec Group so that the two groups constitute a conformable sequence of Cambro-Ordovician age. Additionally, he suggested that the folds that are obvious in the field and which affect the Shickshock and Quebec Groups may be ascribed to the Acadian orogeny.

MacGregor (1962) carried out detailed geochemical and petrological work near Mt. Albert (Fig. 1b). He concluded that the body was a serpentized peridotitic-dunitic intrusion, and he traced a thin, but well defined contact aureole, from this a biotite sample yielded a K-Ar age of 495 ± 35 m.y. (Lowdon et al., 1963), which suggests emplacement in early Ordovician times. Because the schistosity in the surrounding Shickshock metavolcanics swings around the igneous body, MacGregor suggested forceful intrusion postdating the schistosity.

The earlier deformation that produced the schistosity would therefore have taken place no later than the early Ordovician, (see also Sikander and Fyson, 1969).

Recent work in the study area includes paleontological and stratigraphic studies in the Mt. Hog's Back area south of Mt. Albert (Robert,

1966). In the Mt. Richardson area at the southeastern edge of the map area (Fig. 1b), Girard (1967,1969,1971) carried out detailed mapping of a contact aureole (hornfels, skarn) around the Table Top granite. Additionally, Enos (1969), in studying the Middle Ordovician sedimentary rocks (Cloridorme Formation) along the northeastern coast of Gaspé (immediately east of the study area), found chromite and serpentinite fragments in the clastic beds that suggested the presence of pre- or Middle Ordovician ultramafic bodies, such as at Mt. Albert. The uplift and erosion of these igneous bodies would, therefore, be related to an early pulse of the Taconic orogeny.

Lajoie et.al. (1968) in a reconstruction of the stratigraphy and paleogeography of the Matapedia-Témiscouata region (Fig. 1a) recognized an angular unconformity between Cambro-Ordovician rocks of the Quebec Group and Silurian sediments. They also observed that the Taconic folds are much tighter in comparison to the more open or gentle Acadian structures in Silurian and Devonian rocks.

Sikander (1967) and Sikander and Fyson (1969) carried out a detailed structural investigation in western Gaspé from the north shore in Matane area (Fig. 1a) to the head of Chaleur Bay. They recognized two main phases of deformation followed by two less severe phases. The first, which isoclinally folded the Shickshock metavolcanic rocks and part of the adjacent Quebec Group, was contemporaneous with regional metamorphism to greenschist facies. Largely on the basis of the absence of the fold structures from Lower Ordovician rocks to the north, and radiogenic dates such as at Mt. Albert, the deformation was ascribed to an early Ordovician age. The second deformation was considered to have produced the dominant folds in both the Lower Paleozoic and Siluro-Devonian rocks, principally

because these folds are parallel in trend and the axial-plane cleavage appears to form a major cleavage fan that includes rocks of both ages. Thus it appeared, contrary to previous interpretations, (except that of Ollerenshaw, 1967) that northern Gaspé, already deformed in early Ordovician time, was remobilized by the Devonian Acadian orogeny.

For the Port-Daniel area on the south coast of Gaspé (Fig. 1a) Ayrton (1964, 1967) envisaged three main tectonic phases. The first folding, affecting only the unfossiliferous clastic rocks of the stratigraphically lowest Maquereau Group, was referred to a pre-Middle Ordovician "Gaspesian Orogeny". Also an angular unconformity was observed between Middle Ordovician rocks and Silurian formations.

As regards more regional work, Pavlides, et.al. (1968) analyzed the stratigraphic evidence for the Taconic orogeny in the Northern Appalachians. In Gaspé Peninsula they envisaged a series of elongated belts, approximately parallel to the main structural trend, where the Ordovician and Silurian were either unconformable, disconformable or conformable. They suggested in this way that the Taconic orogeny was restricted to domains separated by unaffected areas.

Béland (1969), in an outline of the tectonic evolution of the peninsula, suggested three main tectonic stages, mainly defined by regional unconformities and characterized by particular patterns of sedimentation, magmatism and deformation. The first was terminated by the Taconic orogeny which mainly affected the northern Cambro-Ordovician rocks of the Quebec and Shickshock Groups. The second was climaxed by the Acadian orogeny. This was most intense in the southern part of the peninsula and deformation faded out towards the northern belt,

which was completed unaffected. The third stage was characterized by the deposition of post-orogenic coarse clastics of the Carboniferous cover.

Problems

From the previous work the distribution in time and space of the Taconic and Acadian orogenies remains unsolved. There is general agreement that the northern Cambro-Ordovician belt was deformed during an Ordovician Taconic orogeny and the southern belt during the Devonian Acadian orogeny. However, several workers (McGerrigle, 1950, Poole, 1967, Béland, 1969) considered, mainly because of the presence of unconformities, that the northern belt was unaffected by the Acadian orogeny, whereas from a comparison of fold styles Sikander and Fyson (1969) suggested that the dominant folds throughout the peninsula were Devonian.

The differences in interpretation illustrate the problem of reconciling the tectonic evolution of an area suggested by the study of small-scale structures, with that based on stratigraphic data (regional unconformities, etc.)

It was also not clear whether or not there was more than one orogenic period within the Ordovician. An unconformity between Middle Ordovician rocks and Silurian sediments is well-established in many localities (Pavlides et al., 1968), which indicates that a major tectonic event occurred in the latter part of the Ordovician. However, radiogenic ages from the metamorphic minerals forming the foliation of the Shickshock metavolcanics, and those from the contact aureole of Mt. Albert (Lowden et al., 1963), suggest an Upper Cambrian to Ordovician period for the

regional metamorphism and deformation of the Shickshock Group and the southern part of the Quebec Group. Nevertheless, unequivocal evidence that the rocks affected lie unconformably below unmetamorphosed Lower or Middle Ordovician rocks to the north is lacking.

Also previous work does not indicate if the structures of several deformation phases recognized in the Matane area (Fig. 1a) by Sikander and Fyson (1969) are merely present in a restricted area, or if they are regionally distributed.

Purpose of the Present Study

The present work attempts to determine:

- 1 The nature and sequence of fold generations in the Mt. Albert area.
- 2 The nature of the tectonic boundary within the Quebec Group which separates rocks affected and unaffected by the first phase of deformation.
- 3 The age of deformational events and their connection with the Taconic and Acadian orogenies.
- 4 The age of emplacement of the Mt. Albert ultramafic pluton and its relationship to the deformation and metamorphism of the Shickshock Group.
- 5 Detailed morphologies of structures, areal variations, and possible mechanisms of formation, in particular:
 - i The shape of folds
 - ii The interrelationships between flattening, interlimb angle, layer thickness, and wavelength of folds.
 - iii Cleavage types in sedimentary and low-grade metamorphic rocks.

- iv Degree of chlorite orientation in fine-grained rocks as obtained by X-ray diffraction.
- 6. The tectonic setting of North Gaspé by correlating the structural succession in the study area with that established in other areas. This also demonstrates the limitations inherent in correlating small-scale structures between distant areas.

Acknowledgements

The author is indebted to Dr. W.K. Fyson for his supervision, discussion, and criticism of the work.

Thanks are also due to fellow graduates, and in particular Rao Divi for structural discussions, and C. Kamineni for help in petrological problems. Thanks are extended to Mr. N. Miles of the Soil Research Institute, Department of Agriculture of Canada, and A.S. Wong for skilled guidance with X-ray techniques and interpretation.

The author also wishes to thank Dr. R. Kretz for constructive discussions, Mr. H. Areti of the Physics Department of Ottawa University, and Dr. F.P. Agterberg for advice and help in solving statistical problems, and Dr. M. Shaffiquallah of Carleton University for K-Ar age determinations.

Dr. J. Riva of the University of Laval is thanked for paleontological information, and Dr. C.R. Barnes of the University of Waterloo for study of possible conodont-bearing samples from the Mt. Albert area.

Finally the author wishes to acknowledge financial support by the Geological Survey and National Research Council of Canada through grants to Dr. W.K. Fyson, and of support by the University of Ottawa. E.W. Hearn drafted figures 1 and 2 and took 4 photomicrographs.

II STRATIGRAPHY

Introduction

The rocks outcropping over most of the study area are dated or presumed to be Cambrian and Lower to Middle Ordovician in age (Fig. 2, Table 1). They constitute a thick assemblage of metamorphosed basic volcanic rocks (Shickshock Group) immediately north of Mt. Albert, and further north of shales, greywackes, and argillites with minor limestone (Quebec Group). The assemblage appears to be typical of a mobile environment such as prevails in the early stages of many geosynclines.

To the south of Mt. Albert Silurian and Devonian sedimentary rocks, which are well-dated paleontologically, consist of a platform-like sequence of quartz sandstone, shale, limestone and calcareous siltstones. However, the over-all thickness of almost 6500 m (Béland, 1969) indicates a subsidence much greater than would be expected in a true platform environment.

Cambro-Ordovician Volcanic and Sedimentary Rocks

Shickshock Group

In most outcrops metavolcanic rocks of the Shickshock Group (map-unit 1, Fig. 2) are pale to dark green, foliated to massive. The composition varies from chlorite, epidote, albite and quartz with minor amphibole, in the northern part of the Shickshock range, to hornblende, and plagioclase, with less epidote and chlorite in the south.

TABLE 1 ROCK UNITS

Era System	Period	Rock Unit	Description
Quaternary	Recent & Pleistocene		till, fluvio-glacial deposits
	Middle & Lower	6* Table Top Granite	biotite granite & hornfels (6H)
Devonian	Lower	5 York River F.	sandstone, mudstone, shale & volcanics (5V)
		5 Grand Grève F.	calcareous siltstone, sandstone, coquina and limestone conglomerate
		5 Cap Bon Ami F.	shaly limestone, shale
	Upper	4 St. Léon F.	calcareous siltstone, shale fossiliferous limestone
Silurian	Middle	4 Sayabec F.	dolomitic limestone, minor quartz sandstone
		4 Val Brillant F.	white, grey and red quartz sandstone
	Lower (?)	4 Awantjish F.	shale, minor limestone
		UNCONFORMITY	
Ordovician	Middle	3 Siltstone Unit (Cloridorme F.)	siltstone, shale, sandstone, minor limestone
		21 Mt. Albert Serpentinite	serpentinized dunitic-peridotitic ultramafic body
	Lower	2B Argillite-greywacke Unit	argillite, greywacke, limestone, limestone conglomerate siltstone
		2A Phyllitic-slate Unit	slate, phyllitic slate, limestone, limestone conglomerate quartz sandstone & volcanics (2V)
Ordovician & Cambrian		1 Shickshock Group	metavolcanics, minor meta-sediments

* Numbers 1-6 are map-units used in this study.

The metamorphic grade, as pointed out by Mattinson (1958, 1964), therefore appears to increase from north to south within the greenschist facies.

Associated with the chlorite-hornblende schists are layers of fine sericitic schists that possibly represent volcanic ash beds. Apart from these features, original volcanic layering is not apparent. The layering observed, almost everywhere parallel to the schistosity, is marked by variations in color; it is probably due to metamorphic segregation.

The amount of metasediments associated with volcanic rocks in the study area is less than in other areas in Shickshock rocks (McGerrigle, 1954). However, several lenses of grey calcareous phyllites, red and grey slates, and foliated arkosic sandstones, which interfinger with the metavolcanics, were observed, particularly at the boundary of the Shickshock and Quebec Groups (Fig. 2). The most important belt of sedimentary rocks is near Cascapedia Lake, 15 km west of Mt. Albert, where most of the rock is a quartz sericitic schist associated with a few lenses of conglomerate made up of strained phenoclasts of "albite granite" (McGerrigle, 1954, p. 20).

The age of the Shickshock Group in which no fossils have been found, has long been an object of dispute (see p. 6-7).

Three new K-Ar analyses, carried out on samples of amphibolitic schists and biotite schists collected by the author in the Shickshock rocks east of Mt. Albert, yielded biotite and hornblende ages varying from 440 to 474 m.y. (Fig. 2; see also Chap. IV and Appendix I). The average at 465 m.y. is late Lower Ordovician or early Middle Ordovician. Hence the upper limit of the stratigraphic age of the Shickshock Group

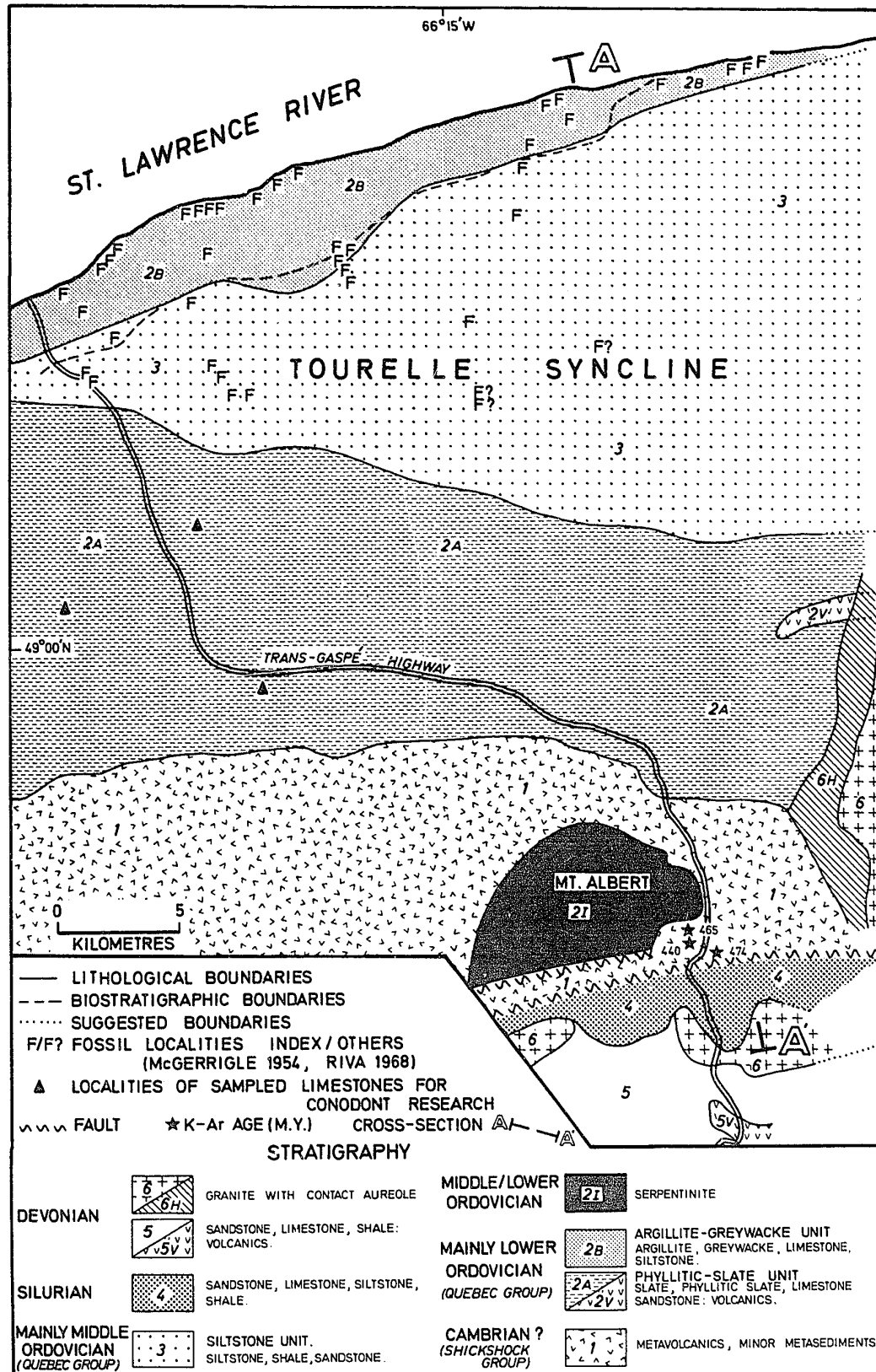


Fig. 2 Lithological map of the Mt. Albert area; modified from previous geological maps (see Fig. 1b). Cross-section A-A' shown in Fig. 3.

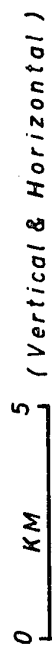


Fig. 3 Schematic cross-section of the Mt. Albert area. Drawn along line A-A' as shown in Fig. 2. All small-scale structures omitted. Map-unit numbers and patterns as in Fig. 2.

may be younger than suggested by radiometric dates obtained previously: notably a K-Ar age of 530 ± 35 m.y. from 50 km southeast of the study area* and an age of 495 ± 35 m.y. from the metamorphic aureole of Mt. Albert (Lowdon et al. 1963). The lower limit depends on the age of the slate sequence immediately to the north and on the stratigraphic relationships between the two units. Neither of these has been definitely established, but a Precambrian age appears unlikely.

Quebec Group

Phyllitic-Slate Unit: This unit (2A) outcrops between the Middle Ordovician rocks, (map unit 3) on the north and the Shickshock Group on the south (Fig. 2), in a belt approximately 12 km wide and 35 km long. The rocks are predominantly phyllitic slates, slates with interbedded limestones, limestone-conglomerates and quartzites (quartz sandstone), with minor volcanics.

Immediately north of the Shickshock Group are phyllitic slates, with a characteristic sheen on the foliation surfaces. They are medium to dark grey in colour, and composed of very fine chlorite, sericite with variable amount of calcite, and quartz. Prominent veinlets of quartz and calcite occupy cleavage and fracture planes.

To the north of the phyllites, black, green and red non-phyllitic slates occur. Apart from the subtle change in metamorphic grade, these rocks are characterized in many places by thin laminae of fine calcareous sandstone or siltstone.

* This age determination is under revision (Dr. R.K. Wanless of the Geochronology Laboratory of G.S.C., person. commun. 1972).

Quartzites (quartz sandstone) of the Kamouraska type (McGerrigle, 1954) are massive, grey medium-grained, locally iron stained and cut by stringers of quartz. These quartzites occur mainly in the eastern part of the unit as discontinuous lenses interbedded with the slates.

Limestones in beds 5-20 cm thick are scattered throughout the area. However, the most prominent outcrops are in the western part of the study area, in elongated, east-west trending, lenses, interfingering with the slates. The limestone is fine to crypto-crystalline, and is interbedded with dark, well-foliated, slates.

Limestone conglomerate is frequently associated with the limestone. Practically all the phenoclasts noted are calcareous, including shaly quartzose limestone. In some localities the phenoclasts are highly elongated and flattened parallel to the bedding and foliation. The matrix is schistose and consists of slate and calcareous sandstone.

Volcanic rocks (map-unit 2V) lie conformably within the phyllitic unit in a restricted belt that extends about 5 km west from Table Top Granite at the eastern border of the study area (Fig. 2). Consisting of basaltic flows with intercalations of tuff and agglomerate (Girard, 1969), the rocks are generally massive, and in comparison to the volcanics of the Shickshock Group they are unmetamorphosed.

The age of these rocks (map-unit 2A, 2V) is not definitely known. On the basis of poorly preserved Middle Cambrian (?) trilobites near the Shickshock rocks in the Matane region (Ollerenshaw, 1967) and general structural conformity with Lower to Middle Ordovician rocks to the north, the rocks are referred to as Cambrian or Lower Ordovician. Because the exact age is very important for structural interpretations, the author

sampled three limestone sequences (Fig. 2) with the hope of finding conodonts. Unfortunately, none of the samples (courteously studied by Dr. C.R. Barnes) yielded fossils.

Argillite-Greywacke Unit: The rocks of this unit (2B) extend inland from the shore of the St. Lawrence River to the axial area of the Tourelle syncline occupied by the Cloridorme Formation (map unit 3) (Fig. 2). Greenish-grey sandstones (greywackes, subgreywackes), dark argillites and shales make up the bulk of the series, with variable proportions of grey, green and red argillites, some limestone, small-pebble conglomerate, greenish-grey chert, calcsiltites, and calcareous sandstone.

The greywackes correspond to the Pillar sandstone of Logan (1846). Graded bedding and lateral variation in grain size are the most prominent features of the depositional units, which are interbedded with argillite and shales. In some sections graded conglomerate with quartz, limestone fragments and feldspar phenoclasts are interbedded with the greywackes.

Many other sedimentary structures were observed in this unit, particularly along the coast. Cross-bedding, convolute laminations, ripple-marks, and slump-type structures appear frequently in calcareous sandstones, calcsiltites and greywackes. Some of these structures were useful in determining the stratigraphic top of the beds; but tops are not consistent, because as McGerrigle (1954, p. 24) observed: "reverse sections may be considered to be quite as common as normal ones in the belt." As discussed later, these rocks are tightly folded so that over-turned and normal limbs of the folds alternate repeatedly.

The age of unit 2B, which is a flysch-like sequence and part of the

Deepkill Formation (McGerrigle, 1954), is fairly well defined by paleontological data as Lower Ordovician (McGerrigle, 1954, Riva, 1968, 1969). The localities where fossils were collected by the authors cited, are shown in Fig. 2.

Siltstone Unit (Cloridorme Formation). Enos (1969) considered all Middle Ordovician rocks on the northern coast of Gaspé as part of a widespread flysch-like, stratigraphic unit, which he named the Cloridorme Formation.

In the study area, Cloridorme rocks, (unit 3) formerly called the Normanskill Formation (McGerrigle, 1954), form the axial area of a broad east-west trending syncline (Tourelle syncline, McGerrigle, 1954), between the argillite-greywackes (unit 2a) and phyllitic-slates (unit 2b) (Fig. 2).

The lithology of this unit is, in part, unlike that described by Enos (1969), and the rocks cannot be classified strictly as a flysch deposit. The main constituents, indeed, are siltstones, shaly and calcareous siltstones, and minor interbedded dark shales and sandstones. Only in two or three localities do greywackes outcrop. The siltstones are in massive beds defined mainly by dark and light colour markers. A well-developed regularly spaced fracture cleavage, can easily be taken for bedding.

The Middle Ordovician age of these rocks is well defined by fossils only in the northern part of the synclinal belt near the argillite-greywacke unit (2b) and in the western portion (Fig. 2). As the paleontologic and lithological boundaries run fairly close together along the northern flank of the syncline, on the southern flank, where fossils are absent, the southern margin of the siltstone unit may also be close to

the boundary between Middle and Lower Ordovician rocks.

Because the rocks outcropping north (argillite-greywacke unit 2B) and south (phyllitic-slate unit 2A) of the Tourelle syncline are lithologically dissimilar, it is suggested that a facies change of coeval rocks occurs beneath the siltstone unit (3) (Fig. 3). Stratigraphically this hypothesis is supported by the fact that in the western part of the study area, where the outcrop of unit 3 narrows, several outcrops of greywackes were observed in the phyllitic-slates, suggesting stratigraphic equivalence to the coastal greywackes.

Further west, beyond the map area, the siltstone unit broadens in outcrop again, so that it is not possible to directly observe a facies gradation between the underlying units 2A and 2B.

Silurian Sedimentary Rocks

The strata of Silurian age are restricted to the southern side of the Shickshock range and in the study area are separated from the Shickshock Group by the Shickshock fault. Hence stratigraphic relations, such as the Taconic unconformity, are not apparent. The strata are highly fossiliferous and well-dated.

Four formations were recognized in the Silurian sequence (McGerrigle, 1954, Robert, 1966), but for the purpose of the present work, all these sedimentary rocks are considered as one unit (4), which forms a normal stratigraphic sequence dipping gently southward.

The oldest Silurian beds (Awantjish Formation) are composed chiefly of green and red shales with interbeds of limestone, calcarenite and

limestone conglomerate. A sequence of white, creamy-grey and red quartzites in massive beds (Val-Brillant Formation) overlies the Awantjish Formation. The upper part of this Formation interfingers with dolomite and dolomitic limestone of the Sayabec Formation.

The St-Léon Formation, the uppermost in the Silurian, is mainly composed of calcareous siltstones, red and green calcareous shales with some intercalations of conglomerates; the rocks are crossed by sills and dykes of acid porphyry.

Devonian Sedimentary and Volcanic Rocks

The basal strata of the Devonian succession (unit 5) form the Bon Ami Formation; it is mainly formed of thinly bedded dark argillaceous limestones. The overlying Grande Grève Formation is marked by coquina layers followed by a thick sequence of silty siliceous limestone, siltstone, and feldspathic sandstone. The youngest formation outcropping in the study area is the York River Formation, consisting of sandstone, interbedded with minor siltstone and a few mudstone and shale beds. Basic volcanics (unit 5V) interbedded with sandstones close to the base of the succession are massive flows, tuffs, and agglomerates.

Igneous Rocks

Mt. Albert Serpentinite

A serpentinized peridotite mass underlying Mt. Albert forms an oval

area approximately 10 km long and 5 km wide (unit 2I). The rock, nearly everywhere massive, weathers to a characteristic buff colour. On the freshly-broken surfaces it is yellowish-green to almost black. The rock is mainly composed of olivine completely or partly altered to serpentine. Locally, crystals of pyroxene are abundant and there are segregations of chromite.

Bordering the serpentinite are dark coloured hornblende-rich rocks, in some places interfingering with garnet-quartz-muscovite schists. These rocks in the contact aureole of the intrusive have yielded a K-Ar age of 495 ± 35 m.y. (Lowdon et al., 1963). The significance of this dating will be discussed in detail in Chapt. III.

Table Top Granite

A small part of the Table Top Granite (unit 6) outcrops at the eastern limit of the study area (Fig. 2). The rock is generally coarse to medium-grained and pink. The amount of quartz is variable (10-30%) and femic minerals, mainly biotite, do not represent more than 20%. Grains appear to be randomly oriented and no alignment was observed even at the near vertical contact with the country rock. Small dark green dioritic masses are enclosed in the granite in a few localities.

Abundant hornfels and skarn form the contact aureole (unit 6H) around the granite body. The hornfels is typically fine grained, hard, massive, and generally rusty on weathered surfaces. Quartz and mica are the main constituents. The skarn is pale green to light grey, generally massive, and it contains mainly quartz, calcite, diopside, garnet and chlorite.

A sample previously collected from the granite yielded a K-Ar age of

420 m.y. (Middle Silurian) (Leech et al., 1963), but two new radiogenic age determinations have given a younger age averaging 350 m.y. (De Römer 1970), (Upper Devonian). This age is in agreement with the cross-cutting nature of the pluton without an internal tectonic fabric, indicating a post-tectonic emplacement. Similar features are apparent in granitic bodies which cut Silurian and Devonian rocks to the south of the study area (Fig. 2).

III STRUCTURAL PHASES

Introduction

Four phases of deformation have been recognized in the study area. According to the spatial distribution of the folds and the linear and planar features associated with them, the rocks in the area lie in three main structural domains (Fig. 5).

Domain 1 includes rocks of the Shickshock Group and of the southern part of the Quebec Group. In addition to later structures, a tectonic foliation lies parallel to the layering and is axial planar to isoclinal folds of the first phase of deformation D_1 . In domain 2, covering the northern part of the Quebec Group, the rocks display widespread micro to mesoscopic folds of the second deformation. Structures of the first deformation are not apparent.

Lastly, the third domain lies to the south of the Shickshock Fault and is underlain by Silurian and Devonian rocks affected only by gentle folds of the Acadian orogeny.

To study the areal variation in orientation of the structures, in particular the D_2 structures, domain 1 and 2 were considered as a whole and subdivided into a variable number of subareas according to the structural features considered (5 subareas for S_2 , L_2 , and 2 subareas for S_3) (see also Fig. 33-38 in Appendix III).

The most prominent features of each tectonic phase are listed in Table 2.

TABLE 2 CHARACTERS OF FOLD GENERATIONS

Deformation Phase	D ₁	D ₂	D ₃	D ₄	
Folds	F ₁	F _{2a}	F _{2b}	F ₃	F ₄
Map Domain	1	1,2	1,2	1,2	1,2,3
Rocks Affected	Shickshock Group (chlorite schists) and Quebec Group (slates)	Quebec Group (slates, sandstone, siltstone, argillites, limestones)	Shickshock and Quebec Groups	Shickshock and Quebec Groups	Silurian, Devonian formations. Shickshock and Quebec Groups
Metamorphism	Regional (Lower to Upper Greenschist Facies)	-	Minor muscovite growth along S ₂	-	Contact to biotite grade
Magmatism	Ultramafic (Mt. Albert) intrusion	-	-	-	Granite emplacement
Distribution	Penetrative	Very Discontinuous	Penetrative	Discontinuous	Discontinuous?
Shape	Isoclinal	Tight	Gentle to Isoclinal	Open, kinks	Gentle warping
Ramsay's Class	1C,2,3	1C,3	1A,1B,1C,2,3	1B,1C,3	
Scale	Micro, few meso	Micro, meso	Micro, meso, few macro	Micro, meso	Macro
Axial Surface Foliation	S ₁ : penetrative slaty cleavage and schistosity	-	S ₂ : slaty, crenulation and fracture cleavage in discrete microspaced zones	S ₃ : local fracture discontinuous cleavage	-
Surface Folded	Jedding (S ₀) volcanic layering (S ₀)	S ₀ ,S ₁ (where present)		S ₀ ,S ₁ ,S ₂	All surfaces
Axial Lineation	L ₁ : trace of S ₀ on S ₁ ; chlorite elongation	-	L ₂ : trace of S ₂ on S ₃ , S ₀ /S ₂ intersection, axes of crenulation in S ₁	L ₃ : S ₀ ,S ₁ ,S ₂ intersections with S ₃ ; axes of kinks in S ₁	-
Orientation	S ₁ : initially near horizontal	-	S ₂ : steep dips SE varying through vertical to NW	S ₃ : vertical NW-SE	Near vertical E-W axial planes
	L ₁ : low plunges E-W	-	L ₂ : plunges NE-SW	L ₃ : steep plunges	
Age of Deformation	465±17 m.y. (K-Ar) early Middle Ordovician (Taconic)	Middle or Upper (Ordovician) (Taconic)	Middle-Upper Ordovician (?)		mid-Devonian (Acadian)

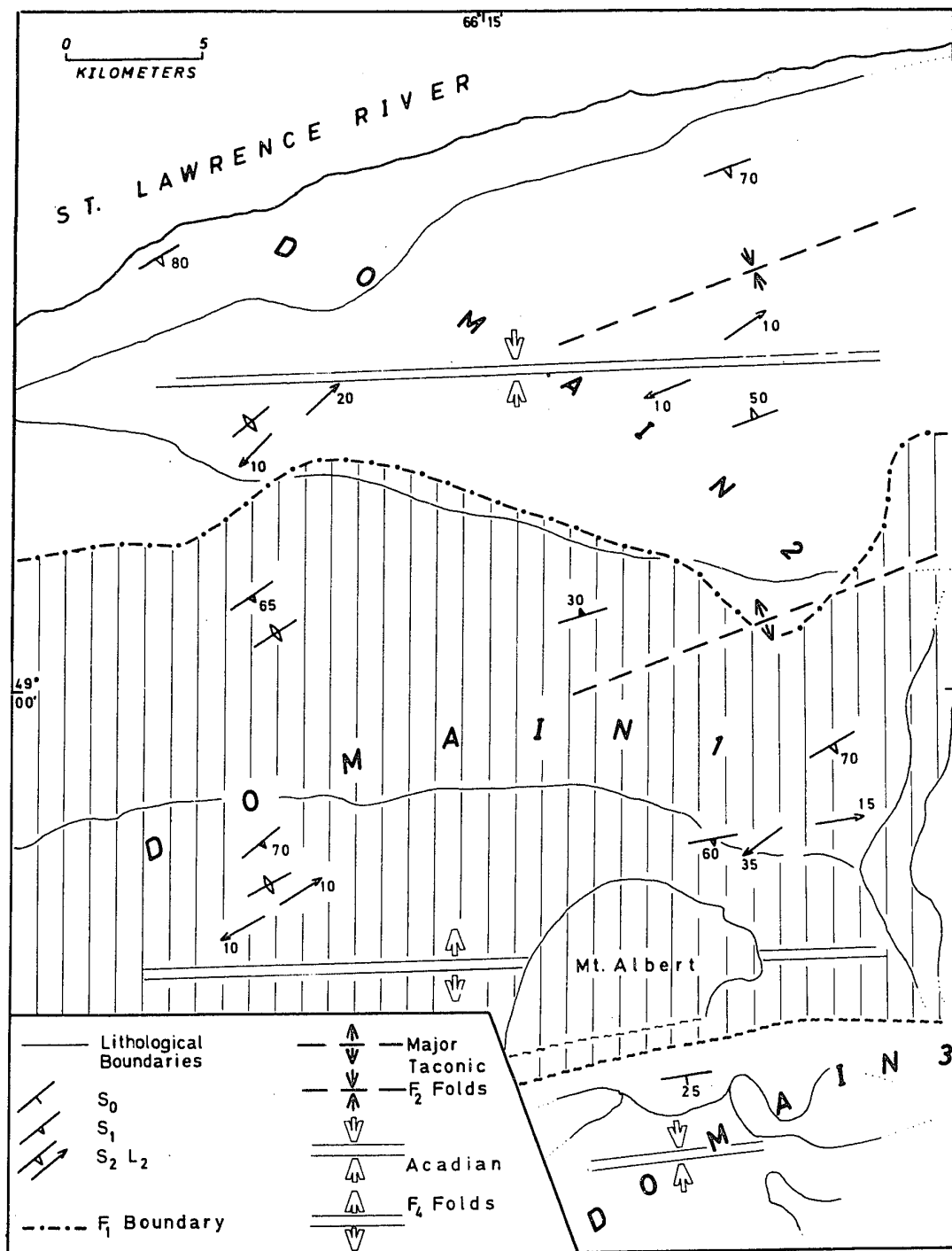


Fig. 5 Generalized structural map of the Mt. Albert area. Domain 1 with F_1 and later folds, domain 2 with F_2 and later folds domain 3 with Acadian folds. Representative structural attitudes shown.

D_1 Deformation

F_1 Folds

During D_1 small-scale intrafolial folds F_1 formed in bedding or volcanic layering, accompanied by development of an axial-surface foliation S_1 . The areal distribution of these structures is restricted to the Shickshock and part of the adjacent Quebec Groups (domain 1, Fig. 4-5). As very few F_1 closures can be recognized in the field or in hand specimen, the information about the character, particularly the orientation, of these early folds is sparse. Nevertheless, in nearly all outcrops the axial-surface foliation S_1 parallels the volcanic layering in the Shickshock Group and the bedding in the phyllitic slates of the Quebec Group. Hence the folds are close to isoclinal. Moreover, because younger folds affecting S_1 plunge at similar angles in domain 1 as those affecting bedding in the adjacent parts of domain 2 where F_1 is absent, presumably before the later deformation S_1 was nearly parallel to the horizontal bedding of domain 2. This suggests that F_1 folds originally were recumbent.

The fold shape varies considerably both throughout the area and in adjacent outcrops (see Chap. V). In Fig. 6 profiles of typical small-scale F_1 folds are traced from thin sections and photographs. Most folds are nearly similar; but in detail their profiles are more commonly those of class 1C in the less ductile, and those of class 3 (Ramsay, 1967) in the more ductile layers of a folded sequence (see also Chap. V).

In the Shickshock metavolcanics, where metamorphism accompanied the

- Fig. 6 F_1 fold profiles. The folds affect bedding planes S_0 , and an axial-plane cleavage or schistosity S_1 is present. Profiles drawn from thin sections.
- a) Magnetite-rich layers (solid black) forming isoclinal folds with limbs parallel to S_1 in chlorite schist; Shickshock Group, (loc. 198).
 - b) Nearly similar F_1 folds in magnetite-rich layers (solid black) in chlorite schist; Shickshock Group. The F_1 structures are crossed by open F_2 folds and S_2 crenulation cleavage (loc. 199).
 - c) Isoclinally folded quartz layer imbedded in pelitic (dashed) matrix, Phyllitic-slate unit, (loc. 51).
 - d) Box fold in imbedded calcite and pelite-rich layers; Phyllitic slate unit. Insert shows relationships to later structures S_2 and L_2 (loc. 253).

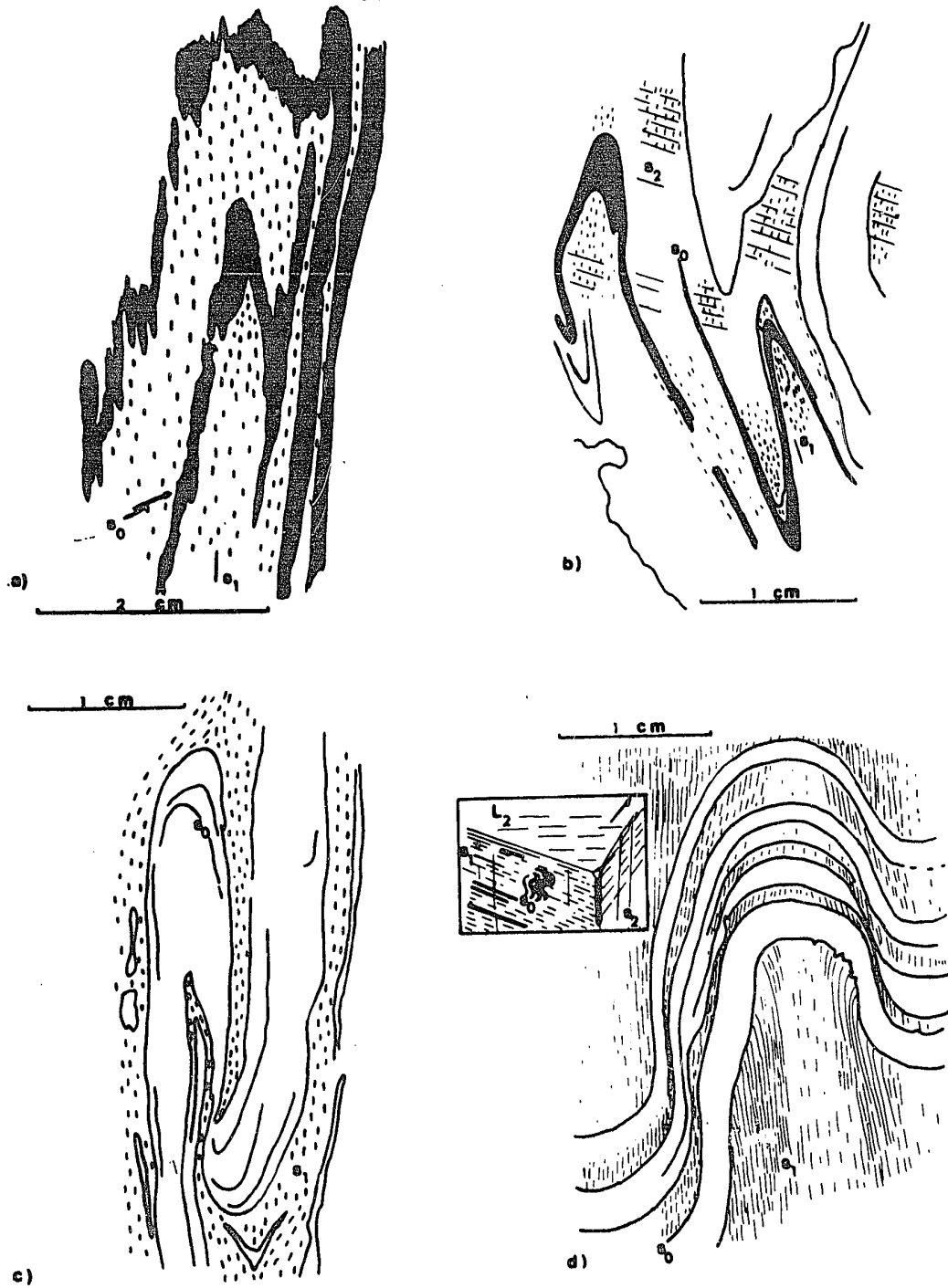


Fig. 6

deformation, the compositional layering is partly or completely transposed along the S_1 schistosity to form isoclinal folds with thickened hinge area (Fig. 6a). Apparently there was little ductility contrast between the layers, and they behaved almost passively. Conversely, in the almost unmetamorphosed slates of the Quebec Group, calcite or quartz-rich beds in the pelitic matrix form folds in which the layer thickness is more nearly uniform (class 1C close to class 1B) (Fig. 6d). The shape indicates that the folding was accomplished by buckling and flattening (Ramsay, 1967).

Small F_1 closures are preserved in the fringe or the "shadow area" of an epidote-quartz knot within chlorite schist (Fig. 7). The knot originally could have been a volcanic bomb or it could have been produced by metamorphic segregation (Mattinson, 1964). Unlike in rocks away from the knot, in the shadow area the layering defined by fine-grained quartz, chlorite, albite, and magnetite crystals is not completely transposed parallel to the S_1 schistosity, hence the F_1 fold closures are preserved.

The shortest dimensions of the knot are perpendicular to S_1 (measured away from the knot). S_1 wraps around the knot, and at two opposite edges is bent to form folds that are coaxial with F_2 folds in the country rock. Several fractures, filled by quartz, are also perpendicular to S_1 .

Apparently during D_1 the knot was relatively rigid and was flattened across S_1 less than the surrounding matrix, hence the warping. Extension along S_1 is indicated by the fractures. Subsequently, the knot was rotated during D_2 deformation and F_2 folds formed at its edges.

Other examples of structures indicating extension along the layering are common in the metavolcanics. Several mafic hornblende-rich layers

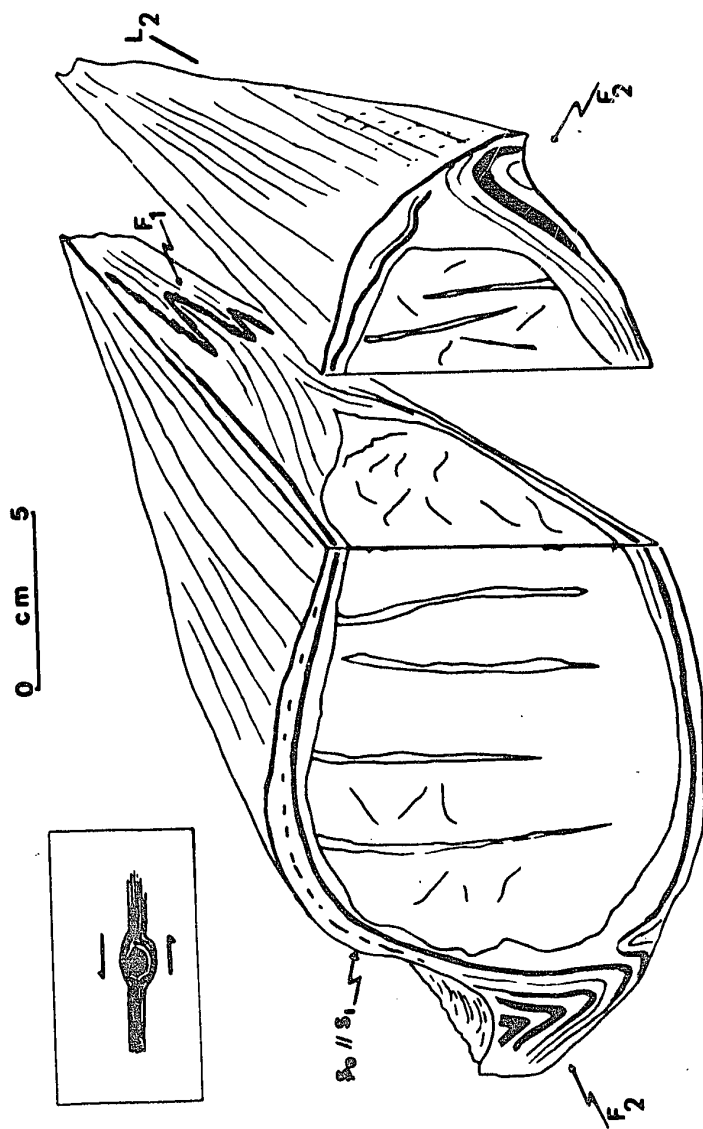


Fig. 7 Epidote knot with adjacent F_1 and F_2 folds in chlorite-epidote-albite-calcite-magnetite layers; Shickshock Group. Knot apparently flattened during D_1 and rotated during D_2 , (loc. 198).

(Plate 1a) show strong boudinage structures. These apparently developed during a late stage of D_1 when the layering was nearly parallel to the axial surfaces of isoclinal folds. D_2 deformation may, however, have enhanced the boudinage process.

L_1 Lineations

Lineations L_1 associated with F_1 folds are few and not easily detectable. A limited number of bedding (S_0) cleavage (S_1) intersections were observed on the cleavage surface in the phyllitic slates as a series of light and dark bands (Plate 2A).

In the Shickshock metavolcanics, apart from a few intersections between layering and schistosity, an apparent L_1 lineation is defined by elongated chlorite porphyroblasts in a medium grained quartz-chlorite-albite-epidote matrix (Plate 1B). The chlorite crystals lie in the foliation plane and define a shallow plunging mineral lineation which is probably parallel to the axes of F_1 folds (Sikander, 1967).

In some outcrops of limestone conglomerate in the Quebec Group, pebbles are flattened in the S_1 foliation surface and elongated, defining a lineation that from its attitude is probably L_1 .

The original trend of the F_1 folds is not obvious due to the intensity of the second deformation. However, the orientation of 40 L_1 lineations, including fold axes, S_0, S_1 intersections and mineral elongations (Plate 1B), suggests that most F_1 folds plunged gently almost east-west at small angles to the trend of F_2 folds (Fig. 33).

For the Matane area, Sikander (1967) indicated that the early (F_1) and later (F_2) folds are coaxial and trend northeast.

S₁ Foliations

The most prominent feature indicating D₁ is not the few F₁ folds observed, but the axial-plane foliation S₁, which occurs as a schistosity in the Shickshock metavolcanic and metasedimentary rocks and a slaty cleavage sensu stricto (see Chap. VI) in the Quebec Group, (domain 1-2).

The schistosity in the metavolcanics is defined by the alignment of syn-D₁ chlorite, hornblende, quartz and epidote; chlorite is warped around feldspar phenocrysts, which are frequently flattened. In the metasedimentary lenses of the Shickshock Group, the schistosity is enhanced by well-oriented recrystallized muscovite. In the phyllites and slates of the Quebec Group, the cleavage, almost invariably parallel to the bedding, is defined by muscovite, chlorite, and flattened quartz or calcite grains that sometimes show strain effects. The nature of this cleavage will be discussed in more detail in Chap. VI.

Frequently it is difficult or impossible to distinguish in the field the S₁ tectonic foliation from a well-developed bedding fissility of sedimentary origin present to the north in the adjacent argillites of the Quebec Group (domain 2). Nevertheless, in many outcrops careful investigation revealed S₀, S₁ intersections of F₁ closures, which are unequivocal proof of a tectonic foliation. Additionally, as crenulations are uncommon in sedimentary fabrics, the presence of a crenulation (produced by the second deformation) on the foliation surface was considered good, but not certain, evidence that the foliation was tectonic.

From the study of thin sections (more than 150) the degree of preferred orientation of platy minerals was qualitatively used as a means

of distinguishing between sedimentary and tectonic fabrics. Also, the presence of strained grains of quartz and calcite was considered evidence for a tectonic origin for the foliation.

Lastly, X-ray diffractometric techniques (see Chap. VII) were applied to evaluate quantitatively the state of preferred orientation of chlorite crystals along the sedimentary and tectonic foliations.

Boundary of F_1 Structures

Using the field, optical, and x-ray criteria a boundary was traced between domain 1 with S_1 parallel to the bedding and domain 2 to the north without S_1 . This boundary was located as exactly as possible in order to understand its nature, an important aspect of the spatial and temporal distribution of the F_1 folds.

A similar boundary within sedimentary rocks north of the Shickshock Group was observed to the west in the Matane area by Sikander (1967), where he envisaged that it represents an unconformity separating older rocks (Cambrian or Lower Ordovician (?)) affected by the early F_1 folds, from Lower or Middle Ordovician (?) rocks (Sikander and Fyson, 1969). However, in the Matane area no other evidence of an unconformity is present, and the supposed age difference across the boundary is poorly defined.

On the basis of the present work, including that extending beyond the Mt. Albert area (Appendix IV), the boundary of F_1 structures, in particular S_1 , is not considered to be an unconformity. The main reasons are:

1. No other structural or stratigraphic features indicating an unconformity

(basal conglomerate, angular discordance, etc.) have been observed along the boundary or in other parts of the Quebec Group.

2. The boundary crosses rocks of identical lithology.
3. Although S_1 cleavage disappears quite abruptly in some localities, at other places it seems to fade out gradually.
4. Although on the southern limb of the Tourelle syncline (Fig. 2) sedimentary rocks of the Cloridorme Formation (map-unit 3), which are upper Lower Ordovician or lower Middle Ordovician, are affected by F_1 structures, these are absent on the north limb in stratigraphically lower rocks (map unit 2b).

Thus rather than lying beneath an unconformity, F_1 structures represent deformation which appears to have been areally restricted within a lower level of the sedimentary and volcanic Cambro-Ordovician pile. In effect, the deformation died out both laterally and vertically.

As S_1 retains its parallelism with the bedding even in virtually unmetamorphosed slates of the Quebec Group near the outer or upper boundary, the dying out of F_1 structures is not accompanied by an opening of the folds. Nevertheless, where S_1 is present in argillites, sedimentary structures, such as ripple-marks, and channels, in parallel beds of siltstone and sandstone are still fairly well preserved, indicating that although folds may be isoclinal the internal strain in some layers was limited.

Age of Deformation

In adjacent areas the age of folds, similar to F_1 , has been considered as Upper Cambrian or Lower Ordovician (Ollerenshaw, 1967,

Sikander and Fyson, 1969), suggesting a correlation with the early disturbance of the Taconic orogeny recorded in southern Quebec (Osberg, 1965, St-Julien, 1967, Béland, 1967). However, because in the Mt. Albert area S_1 cleavage affects rocks of the siltstone unit (map unit 3), which is at least in part early Middle Ordovician in age, this age is a lowest limit to the F_1 folding.

As noted when considering the upper limit to the age of the Shickshock Group, radiometric age determinations yielded values averaging 465 ± 17 m.y. (Lower to Middle Ordovician - see also Chap.IV). The biotites and hornblendes used were aligned along S_1 , so that the radiogenic age also suggests an upper limit to the age of D_1 deformation. Thus as the lower limit determined stratigraphically coincides with the radiogenic limit an early Middle Ordovician age for D_1 is considered probable. This age is similar to that of metamorphic and deformational events elsewhere in the Canadian Appalachians, for example metamorphism and deformation of the Sutton, Bennett and Fleur de Lys belts, and sliding or thrusting of the Klippen in Newfoundland (Rodgers and Neale, 1963, Poole, 1967, Rodgers, 1971).

D_2 Deformation

The folds formed during the second deformation affected all the rocks of the Shickshock and Quebec Groups from the St. Lawrence shore to the Shickshock fault (domains 1,2). These structures do not appear in the Silurian and Devonian formations (domain 3).

Field and laboratory observations suggest that D_2 occurred in two

substages or subphases. The first generated rarely exposed F_{2a} folds that are characterized by a complete lack of axial-plane cleavage. During the second subphase the rocks were intensely folded and various types of axial-plane cleavages formed.

F_{2a} Folds

In only 4 localities could F_{2a} folds be recognized. For example, in calcareous sandstones interbedded with red shales on the coast near Marsoui (Fig. 1b), inclined tight small folds (1-2 m in wavelength) are crossed obliquely by a steeply dipping S_2 cleavage formed during the main D_2 folding. The folds obviously predate the cleavage. In another example in phyllitic slates and thin calcite-rich laminae an F_{2a} micro fold lacking axial plane cleavage is bent by an F_{2b} fold (Fig. 8a). The two folds are exactly coaxial and a crenulation cleavage S_2 , axial planar to F_{2b} , crosses F_{2a} . The F_{2a} fold is distinguished from sedimentary or F_1 structures because it bends bedding and a parallel S_1 . However, for the F_{2a} folds outcropping along the coast, where no tectonic foliation parallels the bedding, a sedimentary origin cannot be rejected.

These F_{2a} discontinuous structures are considered to have formed an early substage of D_2 because they are coaxial or almost coaxial with F_{2b} ; the lack of an axial-plane cleavage suggests non-penetrative deformation.

F_{2b} Folds

Folds grouped as F_{2b} are the most widespread in the Cambro-Ordovician rocks of northern Gaspé. These structures extend from

the eastern tip of the peninsula (Cap des Rosières) to Quebec City and possibly they can be correlated with the main folds affecting Cambro-Ordovician rocks in southern Quebec.

In the study area, F_{2b} folds show a great diversity of style, even in the same outcrop. In profile they vary from isoclinal to open, from almost similar (class 2) to parallel (class 1B) (Fig. 8 a,b,c,d) (see also Chap. V). However, in most cases F_{2b} folds are tight and nearly parallel in the competent layers and nearly similar in the incompetent layers.

The best outcrops of F_{2b} folds are along the St. Lawrence shore (Plates 3-4). Here well-bedded Lower Ordovician sandstones, calcareous sandstones and argillites show countless folds varying in wavelength from a few millimeters to several meters. Examples of disharmonic folding at all scales are frequent (Plate 4). Argillites interbedded with competent sandstones evidently acted readily as planes of décollement.

Disruption, and pinch and swell of the relatively competent sandstones are also common features. It is difficult to establish to what extent such structures are due to tectonic or to sedimentary processes. However, the presence of disrupted F_2 folds closures with boundinaged layers on the limbs and strained bedding fragments suggest that, at least in part, they are due to tectonic processes.

The nature of the folding process during D_2 , apparently largely buckling, is considered in more detail in the section on fold analysis (p. 79).

Due to the lack of recognizable stratigraphic levels and to the poor exposures of the interior, it is difficult to map large-scale F_{2b} folds.

- Fig. 8 Profiles of small-scale F_2 folds. In a, b, and c drawn from thin sections, the folds affect both S_0 and a parallel cleavage S_1 ; an axial plane cleavage S_2 is not penetrative and confined to fine pelite interlayered with calcite and quartz (unshaded).
- a) Isoclinal F_{2a} fold with no axial-plane cleavage, refolded coaxially by F_{2b} ; Phyllitic-slate unit, (loc. 87).
 - b) Open F_{2b} folds with discontinuous S_2 crenulation cleavage; Phyllitic-slate unit (loc. 15).
 - c) Isoclinal F_{2b} fold of large amplitude. Quartz veins (black) fill tension cracks in quartzitic layers (loc. 68).
 - d) S_2 fracture cleavage at high angle to S_0 in the hinge area of mesoscale F_2 fold in massive siltstone of the siltstone unit, (map-unit 3); drawn from photograph, (loc. 203).

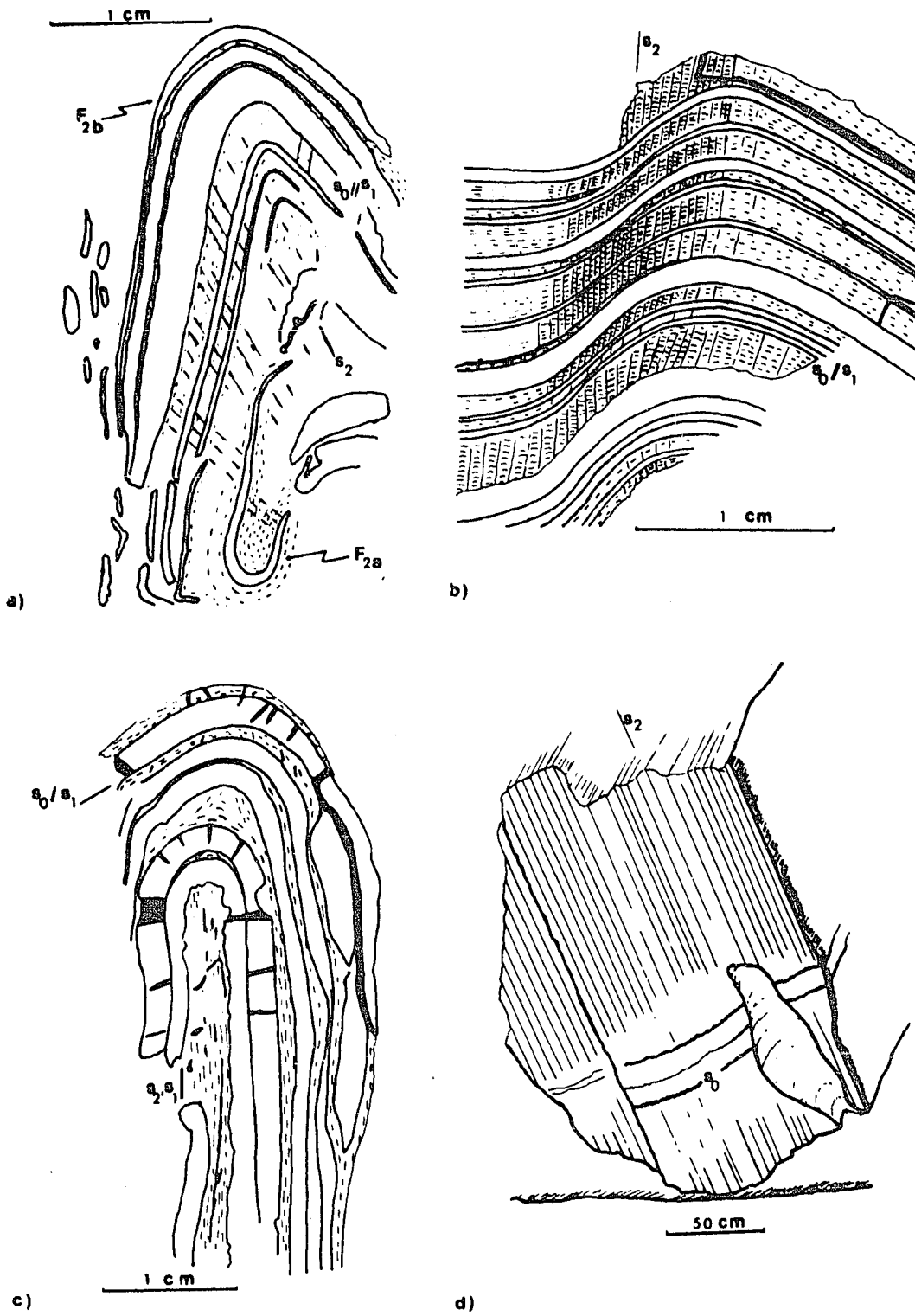


Fig. 8

The present outcrop pattern, as discussed later, is not related to these structures.

S₂ Cleavage

According to the type of lithology, and the presence or absence of a previous tectonic foliation parallel to the bedding, S₂ occurs as a slaty cleavage sensu lato (Chap. VI, p.103) crenulation cleavage or fracture cleavage.

In the sedimentary rocks of the Quebec Group, where S₁ is absent (domain 2) a well-formed fracture and slaty cleavage sensu lato are common, frequently in tight folds at small angles to the bedding. In the slates and particularly in the metavolcanics (domain 1) partings along S₂ cleavage are not always evident, but a well developed crenulation affects most S₁ surfaces. In the metasedimentary rocks of the Shickshock Group, S₂ cleavage is enhanced in some examples by recrystallized muscovite and chlorite aligned along the cleavage surfaces.

In the poorly stratified massive siltstones of map-unit 3, the cleavage, both fracture and "slaty", appears much better developed than in nearby and underlying well-bedded sandstones and argillites (map unit 2b). The same general relationship was observed outside the map area in the Témiscuata region, near Cabano (Fig. 39), both within Lower Ordovician and Devonian rocks.

It is suggested that the lack of a marked stratification, that is of a ductility contrast, delays the folding by buckling, and hence a great part of the strain is achieved by layer parallel shortening (Ramberg, 1964). Flattening, therefore, occurred not only during and after folding, but

also before. Under these conditions cleavage, which is considered to form in the principal plane of strain normal to the plane of shortening (Ramberg, 1963, Ramsay, 1967, Ramsay and Graham, 1970), should develop better than in folds where strain is resolved largely by flexural slip.

L₂ Lineations

Lineations axial to the F_{2b} folds include axes of small scale folds, the trace of bedding (S₀) on S₂ cleavage, and in domain 1, S₁, S₂ intersections. The latter most commonly occur as a well-developed crenulation on the S₁ foliation. No lineations defined by mineral elongations were seen because very little metamorphism (see Chap. IV p. 58) accompanied this deformation.

The F₂ folds are doubly plunging and individual axes and axial lineations curve and plunge at variable angles. As observed for folds of similar form in the Matane area (Sikander, 1967), the plunge changes rapidly, even in the same outcrop, from almost horizontal to nearly vertical.

The curvature of the hinges cannot be ascribed to the interference pattern of fold sets in two directions because, as in the Matane area, no alignment of crestral domes or troughs is apparent, and strongly curved hinges occur side-by-side with those that are almost rectilinear and horizontal. Possibly the curvature resulted from variable extension in the dip direction of the axial surfaces (Sikander and Fyson, 1969), (recent work in Precambrian rocks near Bancroft has demonstrated that the steepness of fold axes may be correlated with the amount of flattening which accompanied or followed the buckling process, Divi, 1972). Thus in

considering the attitudes of L_2 throughout the area (Fig. 33-36, Fig. 4), the variation can be partly due to heterogenous strain, and in part to later deformational events.

Orientation

On the map of Fig. 4 the modal values of S_2 cleavage and L_2 lineations from one or more adjacent outcrops are drawn. Considering representative values from subareas (Fig. 5 and 35), F_{2b} fold axes and axial lineations vary in trend from northeast, in the southwestern part of the area, to east-northeast in the eastern part, with an accompanying easterly swing in the strike of S_2 cleavage.

On average, throughout the western part of the area S_2 cleavage dips steeply southeastward, and it is axial planar to monoclinal folds overturned northwestward. Similar orientations prevail in the Quebec Group of rocks to the west (Sikander, 1967), and further southwest to Rivière du Loup (Fig. 32) where the folds are more overturned.

Near the shore in the eastern part of the study area the mean dip is also southeastward, but between 4 and 6 km to the south the cleavage passes through the vertical and further south dips consistently northwest in a zone up to 15 km wide (Fig. 35, subarea III). Farther south S_2 cleavage again changes dip direction and is inclined to the southeast for approximately 15 km south to the Shickshock fault, (Fig. 35, subarea IV). The changes in dip of the cleavage, which are accompanied by similar changes in dip direction of the surface (S_0 and S_1) folded by F_{2b} , are interpreted as evidence of two major F_{2b} structures, a syncline and an anticline (map and cross-section of Fig. 4). The axes of the major folds

are thus considered as axes of convergence and divergence of the axial-planes and cleavage of small-scale F_{2b} folds. Unfortunately, the stratigraphic evidence is not sufficient to demonstrate or disprove by outcrop patterns the presence of the major structures.

As S_2 , with local exceptions, dips northwest only in the eastern part of the study area, the major F_{2b} folds apparently die out westward.

Strain in Deformed Sandstone Dykes

Near Ste-Anne-des-Monts, several sandstone dykes cross-cut the beds at various angles. The stratified rocks are composed mainly of an alternating sequence of argillites and calcareous sandstones with minor limestone.

The injection of the dykes probably occurred shortly after the sedimentation or in an early stage of the compaction. Therefore, during compaction and possible tectonic overpressure perpendicular to the bedding, these sandstone dykes were folded or stretched according to their initial orientation with respect to the bedding. Dykes making a great angle with the stratification were obviously shortened and folded; those at small angle to the bedding were stretched to form boudinage structures similar to those displayed by veins in metamorphic rocks (de Sitter, 1964, Ramsay, 1967).

A tightly folded sandstone dyke of nearly uniform thickness (Fig. 9) was observed in red argillites bounded by two calcareous sandstone beds. A straight line between the junctions with the bounding beds is nearly at right angles to the bedding suggesting that the dyke originally was also near perpendicular. The perpendicular distance between the two

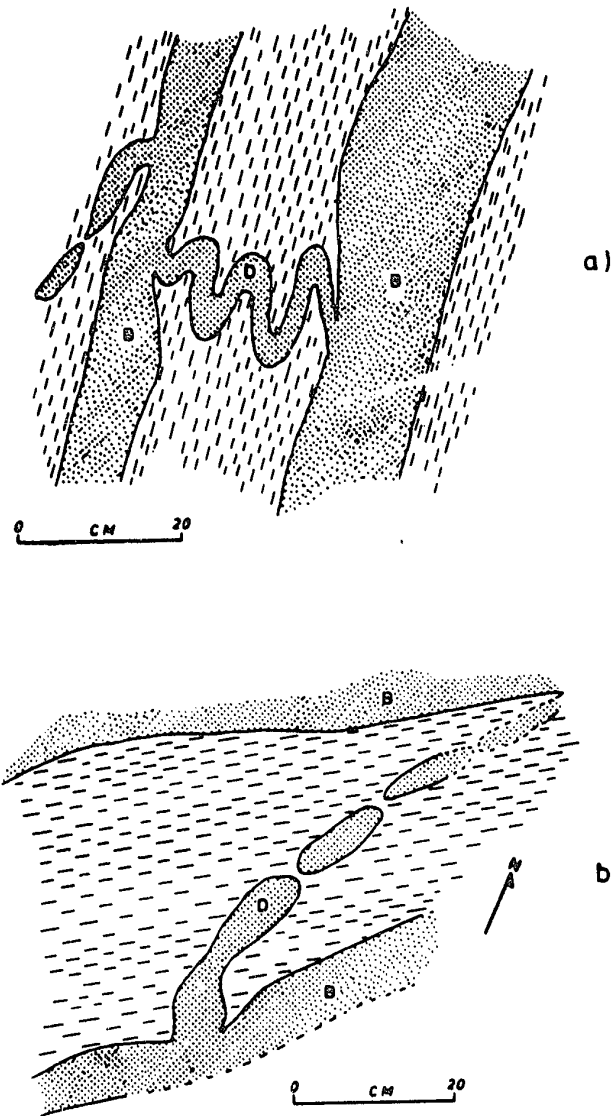


Fig. 9 Strain in deformed sandstone dykes
a) Sandstone dyke (D), between calcareous sandstone beds (B), tightly folded during compaction and tectonic compression; argillite-greywacke unit. Drawn from field sketch, (loc. 25).
b) Sandstone dyke (D), originally oblique to the bedding (B), stretched and disrupted during compaction and tectonic compression. Drawn from photograph, (loc. 25). (Plan view).

sandstone beds is 25 cm, and the length of the folded dyke 75 cm; therefore, the minimum amount of shortening that the dyke and enclosing red argillites underwent, was of the order of 65%.

This value does not take into consideration the layer parallel shortening that the dyke may have undergone before folding, the extra shortening of the more ductile argillites, and the possibility that the dyke was injected after compaction started. Therefore, a value of 70% is probably closer to the total amount of shortening of the argillite layer.

This figure is significantly greater than maximum compaction of less than 50%, in argillites considered to be due to the weight of overburden (Athy, 1930, Hedberg, 1936). It follows that shortening of the order of 20% perpendicular to the bedding was produced by tectonic compression such as would have acted across the limbs during the latter stages of tight F_2 folding.

Age of Deformation

As noted in the introductory chapter, the age of F_{2b} folds, (the dominant folds in the Cambro-Ordovician rocks of North Gaspé) has been in dispute: an Ordovician age (Taconic) is indicated in several localities by tightly folded Cambro-Ordovician rocks lying unconformably beneath Silurian strata.

For example, at Cap des Rosières, at the eastern tip of the Peninsula, (Cumming, 1961), F_2 -type folds underlying gently dipping Silurian rocks are cut by a well-defined erosional surface. This relationship is impossible to explain by the alternate hypothesis of Acadian deformation,

which, largely based on parallel trends of underlying and overlying folds, attempts to explain apparent unconformities elsewhere as the result of disharmonic folding between thin-bedded Ordovician shales and massive Silurian rocks (Sikander and Fyson, 1969).

In the study area the evidence regarding the Acadian age of F_2 folds is not unequivocal, partly because only a small area of Silurian-Devonian rocks was investigated, and partly because these rocks are separated from the Cambro-Ordovician by the Shickshock fault.

Nevertheless, the difference in intensity of folding north and south of the Shickshock fault is striking: folds are tight to the north, open or gentle to the south. This difference could be ascribed to disharmonic folding of rocks of contrasting lithology, as suggested by Sikander and Fyson (1969). But if this was a significant factor during D_2 deformation, the Shickshock metavolcanics, already affected by tight F_1 folding, probably would have behaved even more rigidly than the Silurian sandstones. Also, unlike in the Matane area described by Sikander (1967), in domain 3 no cleavage appears in Silurian and Devonian formations, which include members no less pelitic than those in domains 1 and 2 where S_2 is well-formed, (only discontinuous fractures, part of a tension joint system are present). Therefore, features characteristic of D_2 structures appear to be absent from domain 3; and as described later, the large folds in the Silurian and Devonian rocks of this domain appear to be similar in style to late folds (F_4) in the Cambro-Ordovician rocks.

Thus, following the evidence of unconformities beneath the Silurian from outside the study area (see also Appendix IV), and the restriction of F_2 -type folds to rocks no younger than Middle Ordovician (map unit 3),

F_2 structures are considered to be Middle or Upper Ordovician in age. Hence they are referred to as Taconic structures.

D_3 Deformation

F_3 Folds

D_3 structures which are discontinuous and not everywhere apparent vary in character according to the lithology and the location. In the argillite-greywacke unit (2b) and siltstone unit (3) upright gentle and open rounded folds predominate, whereas in the phyllitic-slates and metavolcanics (domain 1) kinks bands and microkinks are more common.

Orientation

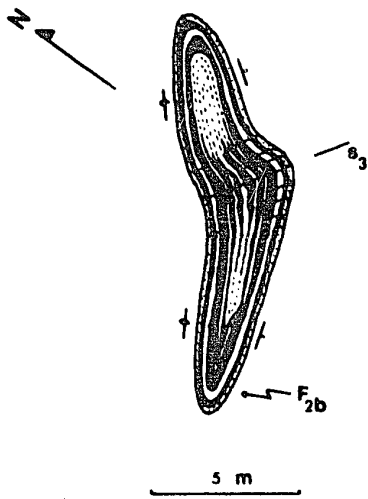
In the excellent exposures along the shore of the St. Lawrence, F_2 folds are refolded or deflected by rounded open F_3 folds, with an axial-planar S_3 fracture cleavage that is subvertical and strikes northwest, (Fig. 10 a,c)

Similar folds were not detected inland probably because of the scarcity of good exposures, but in several outcrops an imperfectly formed and widely spaced fracture cleavage strikes northwesterly and dips steeply like S_3 on the coast.

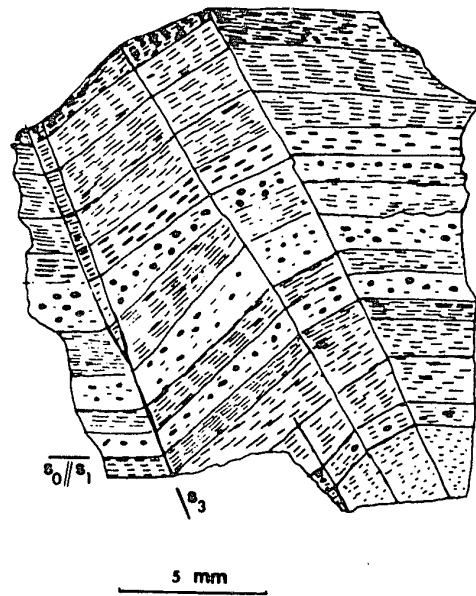
Where S_1 foliation parallels the layering in the metasediments and metavolcanics (domain 1), northwesterly trending, steep plunging, microkinks and kinks appear (Plate 5, Fig. 10, b,d). Frequently the morphology of these structures can be observed only in thin section; in outcrop they appear as a fine crenulation or striation. In the western part of

Fig. 10 F_3 folds.

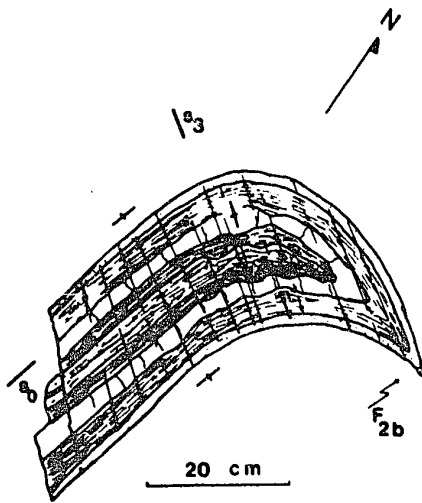
- a) Plan view of double plunging F_2 fold, refolded by open F_3 flexure with vertical axial fractures S_3 striking north-west; argillite-greywacke unit. Field sketch, (loc. 126).
- b) F_3 kinks in thinly layered muscovite-chlorite-quartz metasedimentary rocks; Shickshock Group. Drawn from thin section (loc. 238).
- c) Plan view of F_2 closure refolded by F_3 and crossed by widely spaced vertical S_3 fracture cleavage striking north-west; argillite-greywacke unit. Drawn from photograph, (loc. 136).
- d) Intersecting S_1, S_2, S_3 cleavages in phyllitic slates with thin quartz laminae (dotted). Phyllitic-slate unit. Drawn from thin section, (loc. 410).



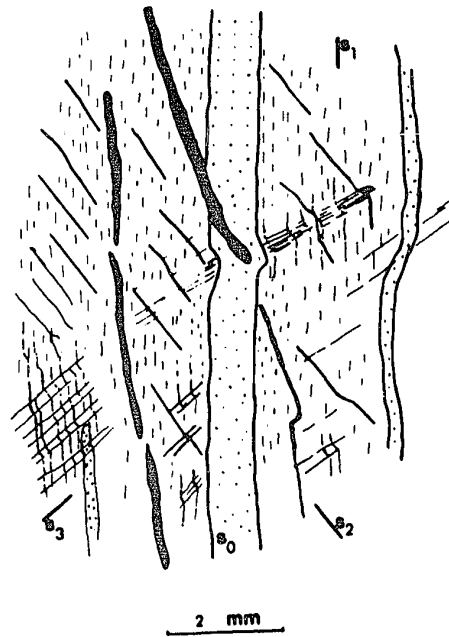
a)



b)



c)



d)

Fig. 10

the metavolcanics, kinks are generally more penetrative and the length of the short limbs increases to several mm. In a few localities small (10-20 centimeters in wavelength) steeply plunging folds with sharp hinges are associated with the kinks and considered to belong to the same generation.

The orientation of the S_3 axial planes of the folds along the coast and of the kinks inland is statistically similar (Fig. 37-38). The differences in character can easily be explained as a response to the presence or absence of S_1 , a well-defined surface of anisotropy. Thus, with no evidence at present of whether or not they mutually interfere, the coastal folds and the kinks are considered to be of the same generation.

However, in the Matane area similar structures were ascribed to different deformational events (Sikander, 1967). Here, open folds near the coast trend east-west, and well-developed conjugate sets of kinks in the Shickshock rocks trend northwest. The folds of the coast were considered to be the result of a sinistral-strike movement along the northeast striking F_2 axial foliation S_2 . The conjugate sets of kinks suggested, conversely, compression along S_2 in the axial direction of F_2 (Sikander and Fyson, 1969).

Age of Deformation

The age of these folds is not well known. Tentatively, because these structures do not appear in Silurian and Devonian rocks, they are ascribed to a last minor deformational event during the Taconic orogeny; in many tectonic belts, indeed, kinks occur during the last "pulse".

D_4 Deformation

F_4 Folds

The last deformation of the investigated area generated large map-scale gentle or open folds trending nearly east-west, which can be traced only on the basis of stratigraphic data. These structures determine the present outcrop pattern in both the Cambro-Ordovician and Siluro-Devonian rocks (domains 1,2,3).

The Tourelle syncline (Fig. 2), the large antiform in Shickshock rocks, and the gentle folds in the Silurian and Devonian rocks are considered to be D_4 structures formed during the Acadian deformation (Fig. 4-5). A difficulty in this interpretation is that no small-scale structures are recognized in the previously folded Cambro-Ordovician rocks which can be ascribed to the D_4 deformation. Essentially the tight earlier F_2 folds mask areal changes in general (sheet) dip that are of the order of merely a few degrees.

The Tourelle syncline did not form during D_2 deformation, because S_2 cleavage obliquely crosses the lithologic boundaries and the axial trace of the syncline at an angle which varies from 10° in the eastern part of the area, to 30° in the western (Fig. 5,36). A similar situation occurs in the Shickshock Group; here the average strike direction of S_2 is at an angle of almost 40° to the northern boundary of the meta-volcanics.

Age of Deformation

The relatively small amplitude and long (approximately 20 km) wave-

length of the Tourelle and Shickshock "warps" (cross-section Fig. 4) are comparable to that of Acadian folds in Silurian and Devonian rocks of central Gaspé (cf. compilation maps of McGerrigle and Skidmore, 1967, and of Sikander and Fyson, 1969). Therefore it is suggested that during the Acadian orogeny a gentle warping also occurred in the Cambro-Ordovician Taconic belt. Nevertheless, the structures forming the outcrop pattern north of the Shickshock fault could also be late Taconic features. This alternative hypothesis which involves a later Acadian deformation producing structures that are parallel in form and orientation to those north of the fault, cannot be rejected, but it is considered improbable.

General Structural Interpretation

The concept of F_4 warping unrelated to small-scale structures as recorded in the field is important in understanding the outcrop patterns throughout northern Gaspé.

In the southern part of the study area (domain 1) the average dip of the layering and S_1 is to the southeast. Similar attitudes prevail along the Shickshock belt to its southwestern extremity, where stratigraphic tops (evidence for which is in the opinion of the author not adequate) also face southward (Ollerenshaw 1967). If the dips are directly related to the outcrop pattern, and if a normal sequence is assumed, then the Shickshock rocks are younger than the northerly adjacent phyllitic slates. However, this interpretation (Mattinson, 1964, Ollerenshaw, 1967) is difficult to reconcile with the higher metamorphic grade of the Shickshock rocks in comparison with the rocks to the north. To overcome this

difficulty, Mattinson, (1964) suggested an inversion of the upper levels by the rising of granitic masses. But, as demonstrated by recent radiometric age determinations (de Rômer, 1970), the granites are Devonian and the metamorphism (see next Chapt. IV) occurred in Ordovician time.

Another suggestion which does not accept the evidence of stratigraphic tops as conclusive is that the Shickshock rocks form the core of a large F_1 recumbent anticline closing northward, (Sikander, 1967, Sikander and Fyson, 1969); the phyllitic slates of the Quebec Group to the north would, therefore, structurally underlie the metavolcanics although they are stratigraphically higher.

In the present work an alternative explanation is suggested (cross-section of Fig. 4). The Shickshock rocks are considered as stratigraphically lower and older than the Quebec Group, and the southeasterly dips of the layering reflect small-scale F_2 folds imposed on previously almost flat-lying bedding, layering, and S_1 foliation. But Acadian D_4 gentle folding warped the enveloping surface of the F_2 folds on such a large scale that these structures are shown only by the map distribution of rock units, and the attitudes of F_2 folds at the outcrop scale remain virtually unaffected. (An exact graphical representation of this structural interpretation is made difficult in the cross-section of Fig. 4, by the fact that F_2 folds are not to scale).

Faults

The Shickshock fault, striking east-northeast, separates the Cambro-

Ordovician rocks of domain 1 from the Silurian and Devonian formations of domain 3 (Fig. 4-5). This major tectonic feature, which extends westward for more than 100 km, is mainly defined by stratigraphic relations and by topography. In a few localities a tectonic breccia is exposed in the fault zone, but exposures in general are poor and the attitude of the zone is not known; however, the relatively straight trace across rugged topography suggests that it is steeply inclined.

The age of the fault is, at least, Middle Devonian (mid-Devonian rocks are truncated in the Matapedia area) and probably late Acadian.

Several small faults, frequently with a strike-slip component, occur throughout the area; in the metavolcanics their strike is generally northeast, in the coastal exposures a few faults striking northwest were observed.

McGerrigle (1954) traced a major fault along the northern boundary of the Shickshock belt. A structural discordance between the metavolcanics and the adjacent phyllitic slates of contrasting mechanical behavior is to be expected, and in some localities the obscure contact does appear marked by a fault. Nevertheless, the stratigraphic evidence for interfingering relations (Chap. II) does not indicate a major displacement between the Shickshock and the Quebec Groups.

Summary and Problems

The structural history of the study area can be synthesized as follows:

- 1) The first phase of deformation (D_1) produced a slaty to schistose folia-

tion that lies parallel to the bedding and to the axial planes of isoclinal recumbent folds. These structures, dated as early Middle Ordovician, die out upward and outward to the north where rocks are unmetamorphosed.

2) The second tectonic event (D_2) generated rarely exposed F_{2a} folds characterized by the lack of an axial-plane cleavage, followed by folds F_{2b} which are inclined to upright, northeasterly trending, and profoundly affect all the Cambro-Ordovician rocks. F_{2b} folds are stratigraphically defined as Middle or Upper Ordovician, and they are considered to represent the main event of the Taconic orogeny.

3) The last recorded Taconic event (D_3) produced sporadic open folds with steeply dipping northwest striking axial surfaces.

4) During the mid-Devonian Acadian orogeny, both Cambro-Ordovician rocks and the Silurian to Devonian formations were gently folded (D_4) on a large scale about nearly east-west trending axes. These structures with accompanying or later faulting, determine the present outcrop pattern.

The above structural history raises several problems which require further investigation.

The distribution in space and time of F_1 folds, that is their dying out upward and laterally, may appear an uncommon structural feature for the northern margin of the Appalachian orogen, where deformational events have been mainly defined and marked by angular unconformities. However, similar hypotheses to account for the spatial and temporal distribution of fold generations have been suggested by other authors (den Tex, 1963, Matte, 1969, Reesor, 1970, Fyson, 1971) in terrains of various ages (Precambrian, Paleozoic, Cenozoic) and in several orogenic belts (Alps, Pyrenees, North American Cordillera). In the cases cited, the spatial boundary of the

folds generally coincides with a significant change in metamorphic grade. In the study area the change in metamorphic grade near the tectonic boundary of F_1 is minimal (see map of Fig. 29), and thus metamorphism does not appear to be a controlling factor in the development of these "lower level" structures.

Another problem which arises from the observation of D_1 structures is the very limited number of F_1 closures and the almost invariable parallelism of bedding layers and S_1 . This situation has been frequently observed in many severely deformed terrains of medium to high metamorphic grade (Turner and Weiss, 1963). Large amounts of strain and extensive transposition of layering, perhaps enhanced by crystal growth could account for these features.

In the study area, however, the parallelism of S_1 and bedding is in virtually unmetamorphosed rocks, and the preservation of sedimentary structures in other layers implies the lack of penetrative internal strain.

IV METAMORPHISM AND MAGMATISM

Introduction

Igneous intrusion and metamorphism occurred during and after the tectonic events which affected the study area. The relationships among deformation, magmatism and metamorphism are briefly discussed with particular emphasis on the age of these events.

Regional Metamorphism

As pointed out by Béland (1957), Mattinson (1964), and Ollerenshaw (1967), the metamorphic grade gradually increases from the St. Lawrence shore to the southern edge of the Shickshock Group. The argillite-greywacke unit along the coast (map unit 2b) is virtually unmetamorphosed, whereas shales and phyllites further south (map unit 2a) display a progressive, although not readily definable, increase in metamorphic grade southward.

In the phyllites with slightly larger crystals than the non-phyllitic slates, the foliation surface (S_1) appears in hand specimen with a luster due to the alignment on the cleavage plane of phyllosilicates (chlorite, sericite, and few small crystals of biotite). Segregation of calcite and quartz layers along S_1 foliation is also apparent in some specimens.

Chlorite-epidote-albite-amphibole schists of the Shickshock meta-volcanics give the appearance of significantly higher metamorphic grade

than the phyllitic slate to the north. However, petrological work and field observation carried out west of the study area by Mattinson (1958, 1964) and Ollerenshaw (1967) indicate that the apparent difference in grade is controlled by the difference in lithology between the meta-volcanics and metasedimentary phyllitic slates. This is also demonstrated in the northern and particularly in the northeastern part of the Shickshock belt, where lenses of phyllitic slates with rare biotite, interfinger with chlorite-albite-epidote-actinolite schists.

The most northerly and least metamorphosed of the Shickshock rocks show a weak fabric characterized by small (about 0.05 mm) grain size, allotriomorphic and often irregular grain outlines, and intermingled minerals. Alteration of the minerals such as plagioclase (mainly albite) to sericite, and amphibole to chlorite is commonly observed.

In many thin sections the mineral assemblage and the fabric appear to be result of complete break down of the Shickshock lavas with little metamorphic reconstitution (Mattinson, 1964). Nevertheless in virtually all the thin sections a careful examination shows some degree of preferred orientation of unbent flaky minerals, parallel to S_1 which indicates at least partial recrystallization.

In considering the apparent contrast in metamorphic grade between the metasedimentary phyllitic slates (map unit 2a) and the adjacent metavolcanics of the Shickshock Group, (map unit 1) the lithological control probably reflects in part the initial variation in chemical composition. Perhaps more significant is the difference in the metamorphic history of the two rock types, as pointed out by Mattinson (1964). The minerals in the metasedimentary rocks were derived from sedimentary assemblages by an increase in temperature and pressure. On

the other hand the initial high temperature low pressure volcanic assemblage, underwent an increase in pressure but a decrease in temperature. It is well known that metamorphic transitions occur more easily during conditions of rising rather than falling temperature, thus high temperature minerals were not completely degraded to low temperature phases.

The rocks show the highest regional metamorphic grade of the study area near the Shickshock fault. Dark green schists contain hornblende prisms up to 2 mm in length, which are optically and dimensionally oriented within S_1 . In a few samples from the area lying between the Shickshock fault and Mt. Albert, small crystals of garnet were observed. However, it is not clear whether this mineral should be ascribed to the regional metamorphism or to the contact effect of the ultramafic intrusive. Garnet is common almost everywhere within the contact aureole of Mt. Albert, but in the Mt. Logan area to the southwest (Fig. 1, area A), where large plutons are absent Mattinson (1964) delineated a garnet isograd in the extreme southern part of the Shickshock.

Metamorphism and Deformation

From the alignment of platy and elongated minerals in S_1 , it appears that crystallization was essentially synkinematic, and thus that metamorphism occurred during D_1 deformation (see also Chap. III).

Minor crystal growth also appears to have occurred during D_2 deformation throughout the area, particularly in the Shickshock rocks and phyllitic slates of the southern part of the Quebec Group. The evidence for the crystal growth is scanty. In the phyllitic slates (map unit 2a)

and to a lesser extent in silty shales within the siltstone unit (3), tiny crystals of chlorite and sericite are aligned along S_2 cleavage. Some of the crystals, however, could be mechanically rotated along the cleavage.

In some localities quartz-rich layers parallel to S_2 alternate with pelitic (chlorite-sericite) layers. In an example from the phyllitic slate (Fig. 13c, Plate 5b), the bedding S_0 is defined by pelitic layers alternating with very thin quartz-rich layers and a parallel S_1 slaty cleavage sensu stricto (see Chap. VI, p. 103), which is defined by aligned chlorite, sericite and tiny strained quartz grains. A secondary cleavage (S_2), which is a type of crenulation cleavage occurs at an angle of 30° to S_0/S_1 . Within the pelitic beds, thin quartz-rich lensoidal layers alternate with the pelitic material which is thus banded. At high magnification (Fig. 13c insert) the quartzitic lenses are seen to be made of strained elongated quartz grains and crenulated mica. Conversely in the adjacent pelitic bands the flaky minerals are undeformed and retain their orientation parallel to S_1 . It is suggested that in an early stage of S_2 development the quartz grains, that were randomly dispersed throughout the pelitic beds, migrated and recrystallized along spaced planes or "domains" which coincided with planes of maximum strain (see also Talbot and Hobbs, 1968). In a more advanced stage of deformation, the quartz crystals were strained and interspersed mica crenulated. Thus in this example the degree of metamorphism was minimal with only quartz showing evidence of migration and recrystallization and at a late stage of D_2 even quartz failed to recrystallize.

Other examples of mineral segregation in slates will be discussed in the chapter on cleavage (VI, p. 105).

In the Shickshock rocks, few examples of metamorphism contemporaneous with D_2 could be detected. S_2 is mainly a crenulation cleavage and only a few unstrained muscovite and chlorite crystals have been observed aligned along the cleavage, mainly in the meta-sediments outcropping within the belt.

In fig. 11 (upper diagram) the time relations of deformation and metamorphism in the Shickshock metavolcanics are tentatively represented. It should be pointed out that the successive deformation episodes (D_1, D_2, D_3) are arbitrarily placed at equal intervals on the horizontal time axis, and the vertical axis representing the temperature is also not to scale. The diagram is simply a graphic way to represent qualitatively the correlation between temperature and deformation. The temperature, indeed, could have fallen between D_1 and D_2 and risen again. The limited evidence suggests that chlorite grew during an early stage of D_2 , but toward the end of D_2 there was little crystal growth even of quartz.

Only tentative comparisons can be made with other areas of Taconic deformation and metamorphism in the Northern Appalachians. However, in southern Quebec the dominant metamorphism also appears contemporaneous with an early isoclinal folding comparable to F_1 , (St. Julien, 1967, Béland, 1967).

Radiometric Ages

An attempt was made to determine the radiogenic age of the meta-

Deformation - Metamorphism

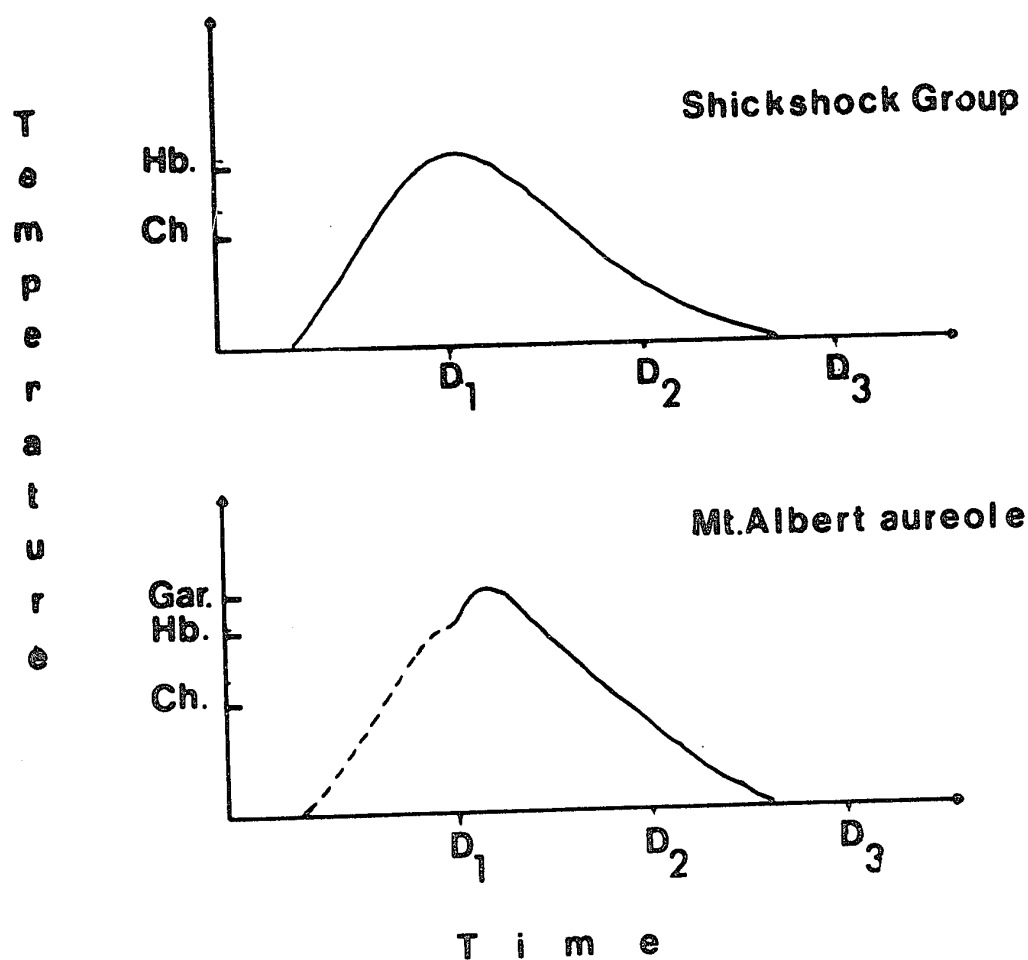


Fig. 11 Time-relations between deformation and metamorphism. Top) Regional metamorphism in the Shickshock Group. Bottom) Mt. Albert contact aureole. Dashed line represents regional and solid line (where ascending) represents combined regional and contact metamorphism.

morphism of the Shickshock Group. Three samples of schist exhibiting S_1 were collected for K-Ar dating from the southern part of the Shickshock Group outside the aureole of Mt. Albert, (Fig. 4). Care was taken to avoid rocks with S_2 cleavage which could enhance the escape of argon. The results are as follows, (see also Appendix I):

From hornblende, plagioclase, K-feldspar schist, hornblende yielded an age of 474 ± 45 m.y. The crystal grains were fresh and well oriented with the long axis within S_1 .

From a sample of biotite, muscovite, plagioclase, and quartz-schist, two analyses were carried out on biotite. The resulting ages 460 ± 25 and 469 ± 26 m.y., vary less than the experimental error.

A third sample (biotite-muscovite-plagioclase-chlorite-hornblende schist) gave an age of 440 ± 24 m.y. This age is apparently younger than the others, possibly due to chloritization of the mica. An average value (taking in consideration the lower reliability of the figure for the chloritized sample) is 465 ± 17 m.y. *, which is late Lower Ordovician or early Middle Ordovician.

* This "weighted" mean was obtained considering the value 440 half as reliable as the others, (that is, the other values were multiplied by 2). The error ± 17 representing the 95 percent of confidence level, was calculated by the standard formula:

$$\left\{ \frac{\sum (x_i - \bar{x})^2}{S_i^2} \right\} / (n-1) \sum \frac{1}{S_i^2} \Bigg\}^{\frac{1}{2}}$$

where

n = number of values; $\bar{x}$ = mean; x_i = individual value;

S_i = error of individual value

As an early Middle Ordovician age is consistent with the field evidence for the age of D_1 (see Chap. III), the dating of about 465 m.y. can be taken to be representative of the age of metamorphism which accompanied D_1 deformation.

Contact Aureole of Mt. Albert Serpentinite

An interesting feature of the Mt. Albert intrusive, which is the largest of the ultramafic bodies of Gaspé Peninsula, is the relatively high-temperature metamorphic aureole, which is unusual in Alpine-type ultramafic intrusives (MacGregor, 1962).

MacGregor (1962) traced the limit of the contact aureole around the body; the aureole varies in thickness from approximately 100 to 500 m. In the southern part of Mt. Albert the lack of an exposed metamorphic aureole suggests a faulted contact.

The thermally metamorphosed country rocks are mainly metavolcanics of the Shickshock Group and in only a few places do metasedimentary lenses outcrop within the aureole. The metavolcanics at the contact with the intrusive are characterized by the following minerals: garnet (almandine-rich), clino-pyroxene, scapolite, biotite, andalusite, oligoclase, epidote, with hornblende as the dominant phase. In the metasediments, quartz, muscovite, and garnet with minor chlorite and biotite were observed. Kyanite also was found (MacGregor, 1962). This mineral, which suggests a higher pressure environment, is uncommon in contact metamorphic rocks. However, as discussed below, the emplacement of the Mt. Albert pluton and the formation of its contact aureole appear to be penecontemporaneous

with D_1 deformation. Thus, deformational pressures might account for the presence of kyanite.

Age of Intrusion

Textures associated with garnet porphyroblasts in the metasedimentary schistose rocks within the contact aureole of Mt. Albert yield information useful in determining the age of the emplacement of the ultramafic body.

Euhedral to lensoidal garnets, (Plate 6) commonly display an internal fabric (S_1) which is defined by the alignment of small elongated, strained quartz grains and minor mica. The boundaries of the quartz are generally corroded, probably as a result of reactions with chlorite to form the host garnet crystal. Some garnets have a poikiloblastic texture due to the large amount of tiny inclusions.

In some garnets the internal fabric is planar and it lies at an angle to the external schistosity S_1 that varies from one crystal to the next, for example in a single thin section the apparent angle varied from 0° to 90° (Fig. 12 c). In other garnets, the internal fabric is slightly sigmoidal (Fig. 12a) and in some examples curves in the outer rim of crystals into parallelism with S_1 .

In addition to the internal fabric, most garnets are crossed by fractures (filled by red-brown iron oxide) which mainly lie perpendicular to the external foliation.

Quartz with minor biotite and chlorite commonly fill "pressure shadows" which in part are asymmetrically disposed about the garnet (Fig. 12b).

The schistosity outside the porphyroblasts is defined by crystals,

- Fig. 12 Garnet crystals in metasediments within Mt. Albert contact aureole. Garnet (black with internal fabric (Si) of aligned quartz inclusions. Matrix grains of quartz (Qtz), and muscovite (M) form S_1 foliation. Drawings of thin sections.
- a) Sigmoidal Si (S_1) in garnet indicating rotation during garnet growth, (loc. 431).
 - b) Garnets with quartz pressure shadows. Internal Si and external S_e quartz fabrics near parallel. S_1 is deflected around the garnets and muscovite crystals are bent, (loc. 516).
 - c) S_1 in garnets at high to low angles to external S_1 (S_e) in the matrix, suggesting variable amounts of rotation, (loc. 516).

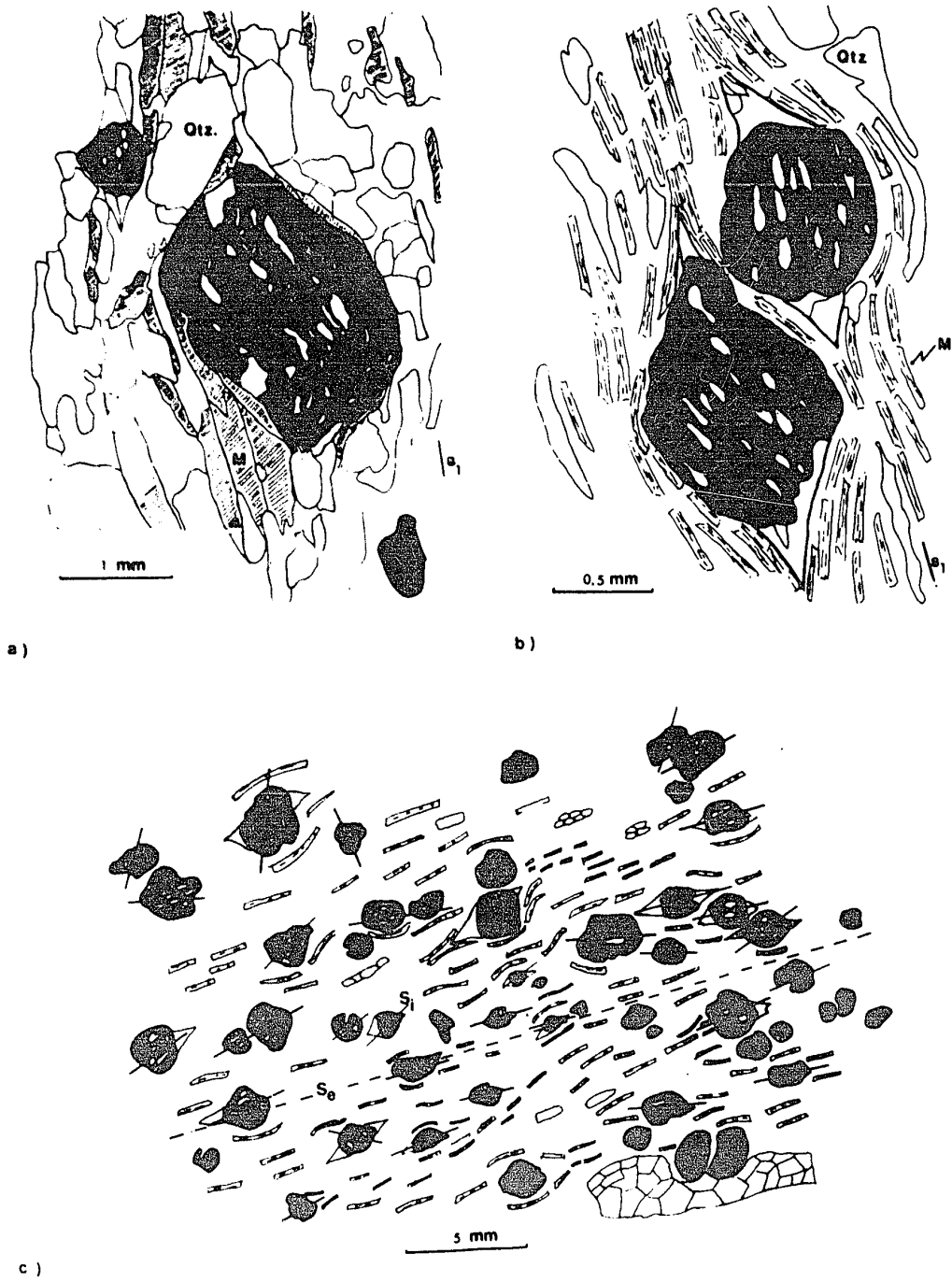


Fig. 12

which are significantly larger than the inclusions, of quartz, muscovite, plagioclase, and minor biotite and chlorite. S_1 is generally deflected around the garnets, and muscovite flakes are bent (Fig. 12b). In some places S_1 abuts against the garnet boundaries.

In the metavolcanic rocks of the contact aureole, garnet also occurs, but included quartz is rare (Fig. 13 a). Quartz is a minor constituent of these rocks and probably it was virtually all consumed during garnet growth. The lack of inclusions and internal fabric make it difficult to ascertain if the porphyroblasts have been rotated.

On the basis of the outlined textural relationships it is apparent that garnet crystals, although confined to the contact aureole of Mt. Albert, did not grow under static thermal conditions. In particular, the sigmoidal internal fabric indicates growth during relative rotation with respect to S_1 (cf. Spry, 1969), and variable rotation is suggested by the variation in angle of the planar internal fabrics to S_1 . Additionally, the coarse texture of the external schistosity as compared with the internal inclusions suggests that the enclosing garnet grew before growth along S_1 was complete.

Considering the regional metamorphism, contemporaneous D_1 deformation and the time of emplacement of the ultramafic pluton, the textures suggest the following relationships.

- 1) During D_1 and the accompanying regional metamorphism of the sediments interfingering with the Shickshock volcanics, muscovite, chlorite, quartz and minor biotite crystallized along the axial-plane foliations S_1 .
- 2) Before the end of D_1 , the dunite-peridotite pluton of Mt. Albert was intruded and garnet formed in adjacent rocks as an effect of the extra heat.

- Fig. 13 Features of crystal growth during contact metamorphism (a,b) and metamorphic differentiation accompanying deformation (c).
- a) Garnet crystals (black) with very little internal fabric in metavolcanics of Mt. Albert contact aureole. In matrix hornblends (Hb) dominant, some diopside (Diop), epidote (Ep), minor quartz (Qtz), (loc. 508).
 - b) Biotite mimetically grown along previous S_1 foliation in hornfels near Table Top granite (loc. 450).
 - c) Fine quartz segregation bands along S_2 crenulation cleavage; Phyllitic-slate unit. Quartz veins (black) along and across S_1 and S_2 , (loc. 430), (See also plate 5b).
- a,b,c drawn from thin sections.

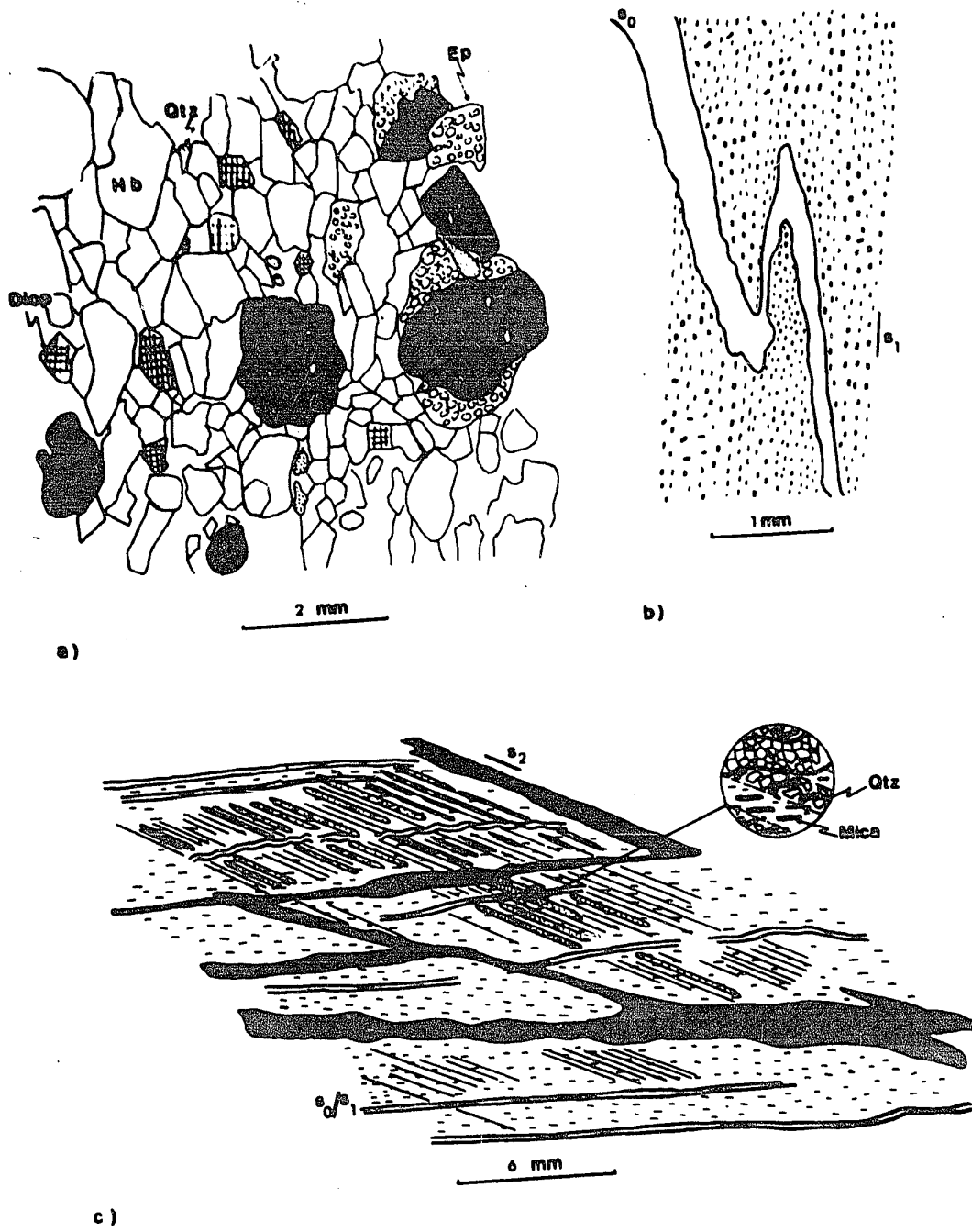


Fig. 13

Because of an excess of SiO_2 , some quartz grains were preserved in the interior of the porphyroblasts.

3) With continued D_1 deformation and metamorphism, perhaps concurrent with further movement of the pluton, flattening and rotation of the S_1 foliation accompanied by differential slip rotated the garnets, both during and after growth. Outside the garnets, crystals such as muscovite and quartz continued to grow along S_1 and a relatively coarse schistosity was formed. Mimetic crystal growth could have continued after D_1 deformation had ceased.

4) At a lower temperature, as the schistosity was deflected around the garnets, muscovite was bent. This was either during the last stage of D_1 , or as would apply if the crystals are mimetic, during D_2 deformation. The garnet was possibly fractured at this time.

It should be also pointed out that the above interpretation is suggested by the field observation that the S_1 schistosity of the Shickshock rocks passes into the contact aureole, and that no extra foliation, except S_1 and S_2 crenulation, was observed within the aureole. Hence, the fabrics in the garnet schists are not produced by the emplacement of the pluton.

It follows that the deflection of the schistosity in the Shickshock rocks around Mt. Albert is not the effect of forceful intrusion of the igneous body into an earlier foliation, as suggested by MacGregor (1962). Rather, late D_1 or D_2 deformation bent S_1 around the pluton in the same manner as foliation was deflected around early kinematic porphyroblasts such as the garnets.

In summary, the regional metamorphism, the intrusion of the ultra-

mafic body and D_1 deformation were penecontemporaneous events (Fig. 11 lower diagram). Therefore following the stratigraphic and radiogenic evidence for the age of the deformation (Chap. III), and contrary to previous interpretations (MacGregor, 1962, Poole, 1967, Béland, 1969), the intrusion appears to have occurred in early Middle Ordovician time. Due to the large experimental error, the K-Ar age of 490 ± 35 m.y. obtained from micas in the aureole (MacGregor, 1962) is not inconsistent with this interpretation.

Contact Aureole of Table Top Granite

Within the 1-2 km wide contact aureole of the Table Top granite, hornfels composed mainly of fine-grained quartz, biotite, and minor chlorite, have a well-developed fabric, which is uncommon in thermally metamorphosed rocks (Spry, 1969). The biotite is undoubtedly a product of contact metamorphism as it disappears gradually towards the margin of the aureole. Nevertheless, the great majority of biotite flakes are aligned along S_1 foliation and parallel layering (bedding), and only a few are randomly oriented (Fig. 13b).

Evidently the effect of the thermal metamorphism has been to enhance the previous S_1 fabric by mimetic crystallization.

It is of note that without the field evidence that the rocks are within a contact aureole, this fabric could easily be taken to indicate synkinematic crystallization.

V FOLD ANALYSIS

Introduction

In the last decade laboratory experiments and mathematical treatments have contributed greatly to the understanding of folding processes. The effect of viscosity contrast, layer thickness, number of layers, strain and stress rate and magnitude have been studied, and a series of equations, which correlate these factors, have been found (Biot, 1957, 1961, 1964, Ramberg, 1959, 1960, 1961, 1963, 1964, Ramberg and Strömberg, 1971, Ramsay, 1962, 1967, Hobbs, 1971).

Less conspicuous is the amount of work done on natural folds. Because the physical properties of rocks during folding are not yet well known, stress conditions are virtually unobtainable, and due to the lack of suitable markers, strain in nature is rarely measurable. Also folded rocks frequently have so complicated a structural history that it is difficult to separate the effect of the different tectonic phases.

Nevertheless, for relatively simple natural folds of various sizes from several terrains, Currie, et al. (1962) found a linear correlation between dominant layer thickness and fold wavelength. The wavelength tended to be 27 times the layer thickness.

These fold measurements by Currie et al. (1962) appear to accord well with the Biot (1957, 1961) theory of the folding of viscoelastic stratified media (according to the Biot theory, the wavelength-thickness ratio is proportional to the cubic root of the ratio of viscosities of

the folded layer and embedding medium. Also laboratory experiments by Biot, et al. (1961) agree with the Biot theory.

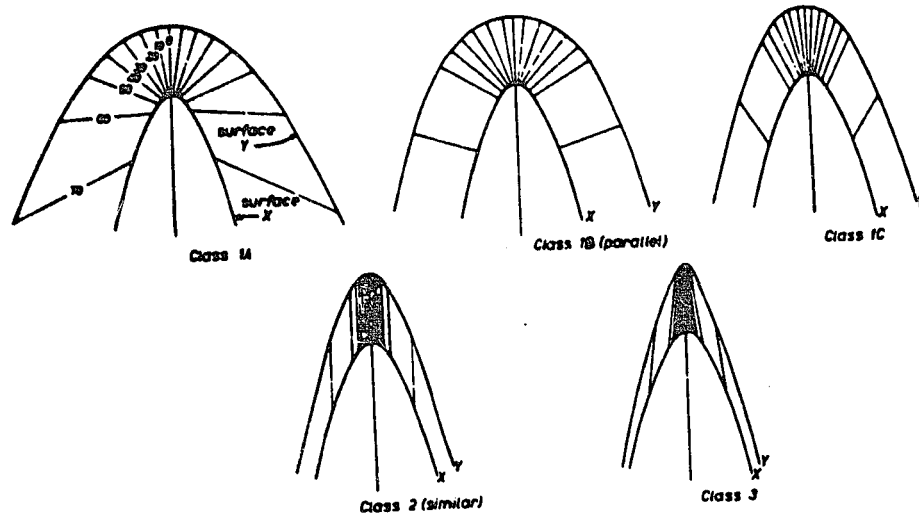
However, other experiments (McBirney and Best, 1962) and investigations of small-scale single-layer natural folds (Sherwin and Chapple, 1968) seem to indicate a large discrepancy between the Biot equation and the experimental and field observations.

Several genetic classifications of fold shapes (de Sitter, 1964, Ramberg, 1963, Donath and Parker, 1964) and non-genetic classifications (Loudon, 1964, Chapple, 1964, Ramsay, 1967, Stabler, 1968), have been suggested. The first are based on mechanical processes which are thought to govern the folding development; the second on morphological characteristics, mainly the profile shape as measured in a section perpendicular to the fold axis.

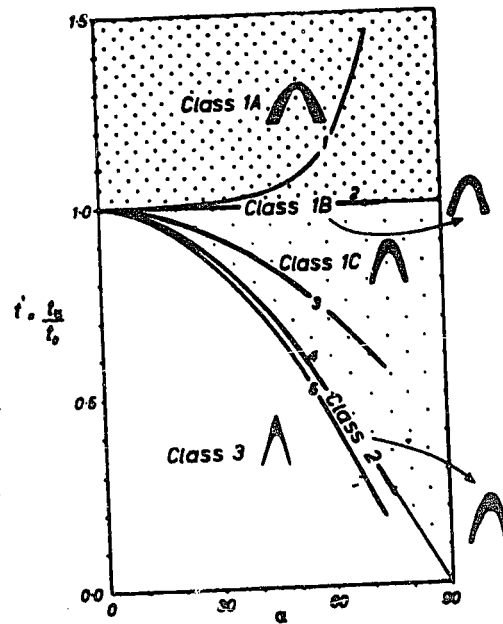
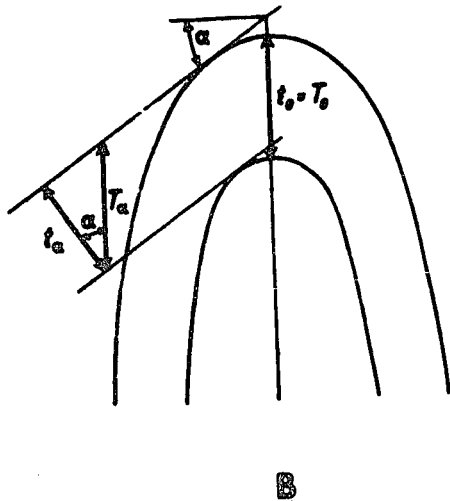
In the present investigation the shape of fold profiles and related aspects are classified on a non-genetic basis. The amount of homogeneous strain (flattening) which affected the layers during folding is estimated from the shape of the folds and compared with the interlimb angle, layer thickness and wavelength. Also the relationship between wavelength and layer thickness is examined, and for few folds the apparent ratio in layer viscosity determined. Lastly, the interlimb angles and estimates of flattening of F_1 and F_2 folds are compared to see if they differ significantly according to generation or areal distribution.

Procedure

Folds in the study area were morphologically grouped into classes 1A, 1B, 1C, 2 and 3 of Ramsay (1967) (Fig. 14). With the limitations



A Fundamental types of fold classes. Dip isogons have been drawn at 10° intervals from the lower to the upper surfaces X and Y .



C Position of t_a plots for the various fundamental fold classes.

Fig. 14 Classification and graphical representation of fold shapes in profile (Ramsay, 1967).

that will be discussed, the classification was useful in the field for a general grouping, and in the laboratory (polished surfaces, thin sections) for more detailed measurements. Each layer was classified separately, and in some cases segments of the same layer within a single fold fell into different classes. How the different classes alternate with each other from one layer to the next enables one to infer something about the processes which occurred during folding.

Ramsay distinguishes the three fundamental fold classes by the change in curvature from one surface to the next: in class 1 the curvature increases towards the interior of the fold; in class 2 (similar folds) the curvature is constant; in class 3 the curvature decreases. The convergence, parallelism or divergence of dip isogons illustrate changes from one class to the next in a series of layers (Fig. 14A). Additionally, class 1 is subdivided according to changes in orthogonal thickness (t): in class 1A, t at the hinge is thinner than on the limbs; in class 1B (parallel folds) t is constant; in class 1C, t is thickest at the hinge.

The variation in curvature and orthogonal thickness are useful for describing folds in the field. Nevertheless it was found almost impossible to distinguish a true parallel fold (class 1B) or a true similar fold (class 2) from the folds of classes 1A, 1C or 3, that closely approach them in shape. In such cases only laboratory analyses allowed an exact classification.

For this purpose, following Ramsay (1967), a convenient method is to plot a graph of α versus t' (Fig. 14B), where $90 - \alpha$ is the angle between the tangent to the folded surface and the axial plane of the fold,

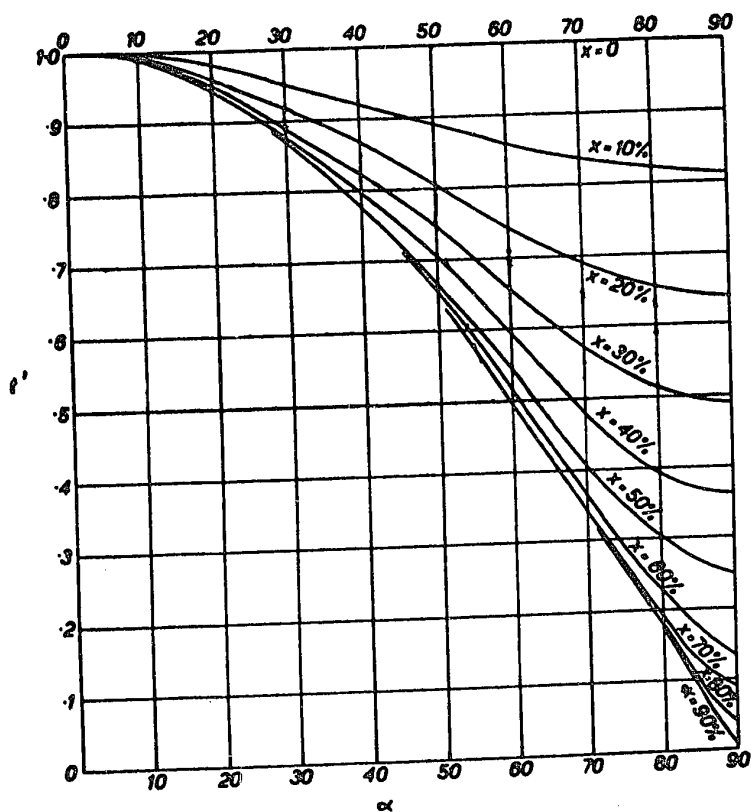


Fig. 15 Graph showing relationships between layer thickness variations and flattening strain in fold profiles.
 t' = ratio of thickness of layer at dip α (t_α) to thickness of layer at hinge (t_0). x = percentage of flattening (Ramsay, 1962).

and where $t' = t_{\alpha} / t_0$; t_{α} is the orthogonal thickness at the point where the tangent is drawn, and t_0 is the thickness at the hinge of the fold. The graph so obtained defines in detail the variations in thickness of the folded layer. In considering folding mechanism, the graph was used to calculate for 1C folds the amount of flattening that accompanied or followed the buckling progress, which is assumed to have taken place during the folding (Fig. 15).

The morphology of a fold is not completely defined with the use of the above classification because fold features like interlimb angle, amplitude, wavelength, and sharpness of the hinge area are not considered. Some of these parameters are taken into account in other morphological classifications (Chapple, 1968, Stabler, 1968), but none of them is suitable for application in the field.

In the present study, the wavelength and interlimb angles were measured in the field and laboratory for a sufficient number of folds to warrant statistical comparisons with layer thickness and the estimates of flattening, as well as to test for areal variations.

Field Classification

Ramsay's classification was used to describe in the field about 300 folds, each affecting several layers and ranging from a few millimeters to several meters in wavelength. Most are F_2 , about 230-240; the remainder includes about 25-30 F_1 and 30-40 F_3 folds.

The result of this field investigation of the fold shape is only semiquantitative, nevertheless it fits well with the limited number of

more detailed laboratory observations.

All folds observed in the field in different lithologies, and of different age, can be grouped in the following way:

Class 1A: very few (1-2%) of the folds can be ascribed to this class and in some cases it is not sure if the thinning of the hinge area and the thickening of the limbs has a tectonic or simply a sedimentary origin. These folds have been observed mainly in limestone beds of the Quebec Group and all are F_2 folds.

Class 1B: a few (5-8%) F_2 folds are true parallel folds. This class is confined to limestone and sandstone layers, which obviously were more competent (less ductile) than the interbedded pelites.

Class 1C: the "flattened parallel folds" are the most common type (about 50%) for both F_1 and F_2 folds in competent layers in metamorphic and sedimentary rocks throughout the area.

Class 2: very few (1-2%) true similar folds were observed. This class is represented by rare (4-5) F_1 and approximately an equal number of F_2 folds in the Shickshock rocks and in the phyllitic-slate belt (map unit 2a).

Class 3: almost all the pelitic incompetent layers (both F_1 and F_2) belong to this class, although without careful measurements they can readily be confused with class 2 folds.

Laboratory Measurements

Among the 300 folds a restricted number (34, of which 7 are F_1 and 27 are F_2) was selected for laboratory work; for this group the exact position in Ramsay's classification was determined for each of a total of 105 layers, mainly calcite, quartz-rich or pelitic (Table 5, p. 185) coming from various parts of the study area.

Fold Classification and Lithology

For the total number of layers measured, the class with the highest frequency is 1C (Fig. 16a). This is in agreement with the field observations. Nevertheless, because more competent than incompetent layers were measured, a better representation of class distribution is given in Fig. 16b, where the layers are divided into two groups: competent and incompetent, and where the relative frequencies are calculated separately.

The criteria used to group the layers as competent and incompetent are qualitative and relative. For example, in a folded sequence of sedimentary rocks with calcite or quartz-rich layers alternating with pelitic layers, the latter are considered incompetent. In metamorphic rocks the more mafic are regarded competent in comparison with more silicic volcanic layers as the mafic layers exhibit less pronounced changes in thickness.

Of the folds in competent layers (Fig. 16b) 65 (88%) belong to the 1C class and only 9 belong to the other classes. It is of interest to note that a few layers (2) form class 3 folds, which are more typical in a sequence of alternating lithologies of the less ductile (competent)

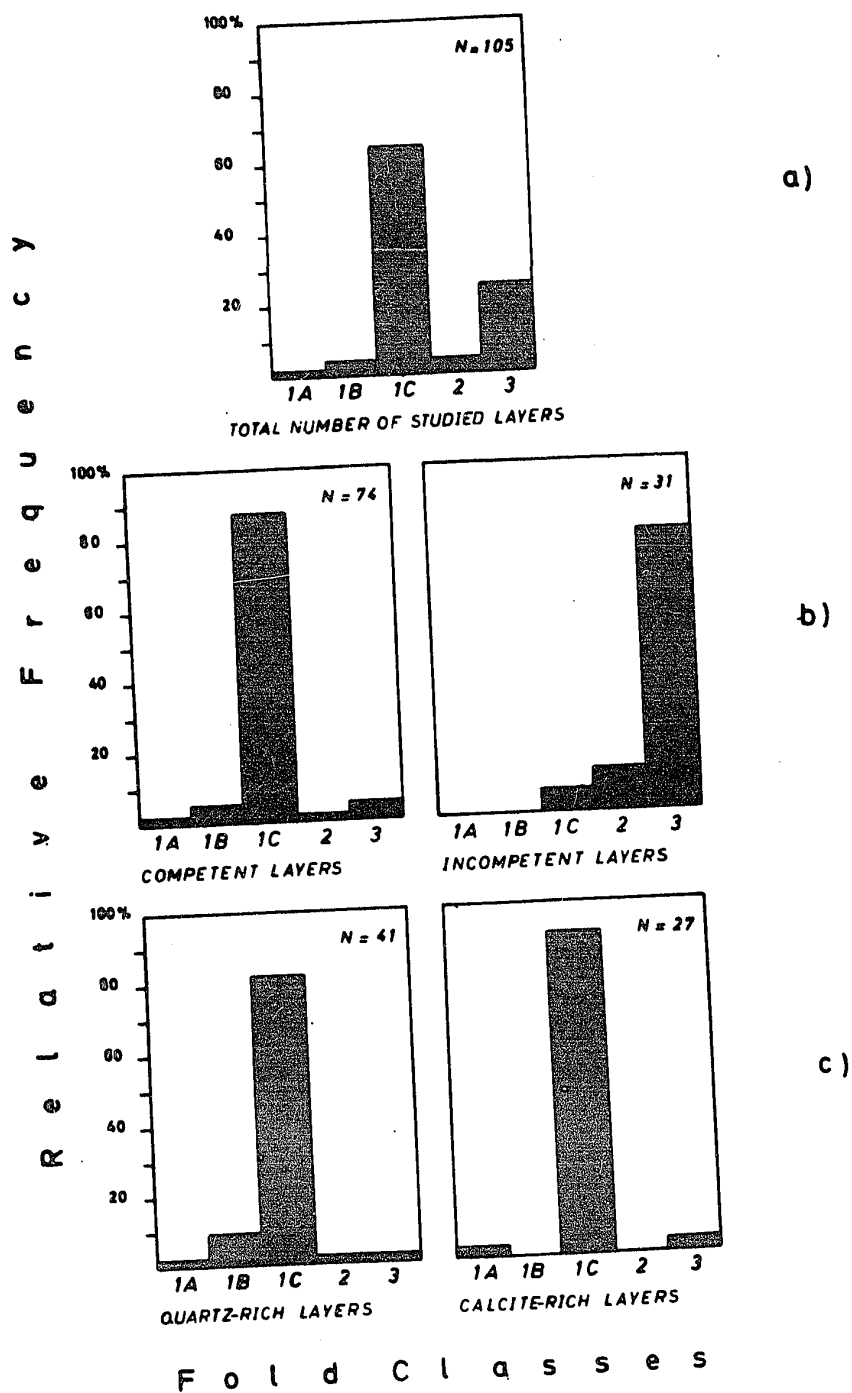


Fig. 16 Relative frequency of Ramsay's fold classes observed in folded layers of the study area. N= number of layers measured.

a) Total number of layers studied in 7 F_1 and 27 F_2 folds.

b) Class frequencies in competent (quartz or calcite-rich) layers, and in incompetent (pelite-rich) folded layers.

c) Class frequencies in quartz-rich and in calcite-rich folded layers within pelitic matrix.

members, (Ramsay, 1967). Thus, in some cases calcite and quartz layers interbedded within pelites appear to have behaved unexpectedly as relatively more ductile members.

For the 31 incompetent layers (Fig. 16b) the modal class is class 3 (80%), and 4 (12%) are class 2 similar folds. Additionally 2 pelitic layers form "anomalous" 1C folds, typical of the competent members.

It appears, therefore, that during folding some members were more or less ductile than expected on the basis of lithology. Probably in the folded sequences, layers of the same composition reacted as competent or incompetent members according to their position in the sequence and their thickness.

In Fig. 16c histograms of fold classes in relatively competent quartz and calcite-rich layers, both interbedded in pelites, are plotted separately to see if there is any significant difference in fold shape. The only appreciable difference appears to be the lack of true parallel folds (1B) in the calcite layers; but in the quartz-rich layers there are only a few such folds (4 out of 41), and a similar low frequency in the 27 calcite layers measured could lead to an absence that was random.

General Fold Mechanism

From the fold classes in the study area it is possible to infer something about the mechanism of folding.

The few 1A folds observed could have been produced either by bending (Ramberg, 1963) or by buckling in the outer fold arc of a layered sequence (Ramsay, 1967), and the few true parallel folds (1B) by buckling, with

simple shear or variable longitudinal strain parallel to the layer boundary.

Class 1C, which is by far the most common type in competent members in both F_1 and F_2 folds, is produced by buckling associated with a component of homogeneous strain or flattening, which can occur during or after buckling. For each strain increment the resultant deformation is a contraction in a direction parallel to that of the principal compressive stress and an extension at right angles (Ramsay, 1962, p.313).

The few similar folds (class 2) observed in the study area may have been formed in part by a shear mechanism, with transmission of differential slip along planes that extend from adjacent layers that are buckled (Ramsay, 1967, Hobbs, 1971). Their association with 1B and 1C folds in the same outcrop also suggests flattening.

The very common type of class 3 F_1 and F_2 folds in incompetent layers apparently is the result of extreme thickening and thinning due to the passive accommodation of highly ductile material between two more competent layers folded by buckling and flattening (Ramsay, 1967).

On the basis of the morphology of the folds, that is the dominance of classes 1C and 3, it appears that the main folding mechanism which operated during F_2 deformation was buckling associated with flattening. The same mechanism may have operated to produce the F_1 class 1C folds, but with these it is possible that some of the flattening occurred during F_2 folding.

Wavelength and Layer Thickness

The concept of "dominant member" was first introduced by Currie,

Patnode, and Trump (1962). The dominant member is the thickest competent member within a sequence of layers folded harmonically by buckling; the dominant member controls the wavelength of the adjacent strata of the sequence. If a competent layer is bounded by two thick incompetent layers, it is considered the dominant member of a lithic unit.

In Fig. 17a the relative frequencies are shown for the ratios of the quarter-wavelength $(W/4)^*$ (Stabler, 1968) to the thickness of 32 layers in 13 folds from separate localities. The ratio $(W/4)/t$ for all 32 layers varies from about 1 to 10 with a mean value of 2.7. The spread of the ratios for 21 dominant members, all that could be distinguished in the limited exposures (Fig. 17b) is nearly half that of all the layers and the mean value of 1.9 is smaller. It is of interest that the range of $(W/4)/t$ ratios and their means are relatively close to values $\{(W/4)/t=2-3\}$ observed in folds of comparable size and in similar lithologies from other areas (Ramberg and Ghosh, 1968, Sherwin and Chapple, 1968).

The quarter-wavelength and mean thickness of the dominant members were compared graphically on log-log paper (Fig. 18). The mean thickness was obtained by averaging the orthogonal thickness measured around a fold, and the quarter-wavelength from the perpendicular distance between the axial surface and the inflection point (Stabler, 1968, p. 344).

On the graph of Fig. 18 a best fitting line was drawn by means of the

* A quarter of wave has been used instead of the whole wavelength, because in folds of irregular shape or in samples where only part of the fold could be observed, this measurement is easier and more reliable.

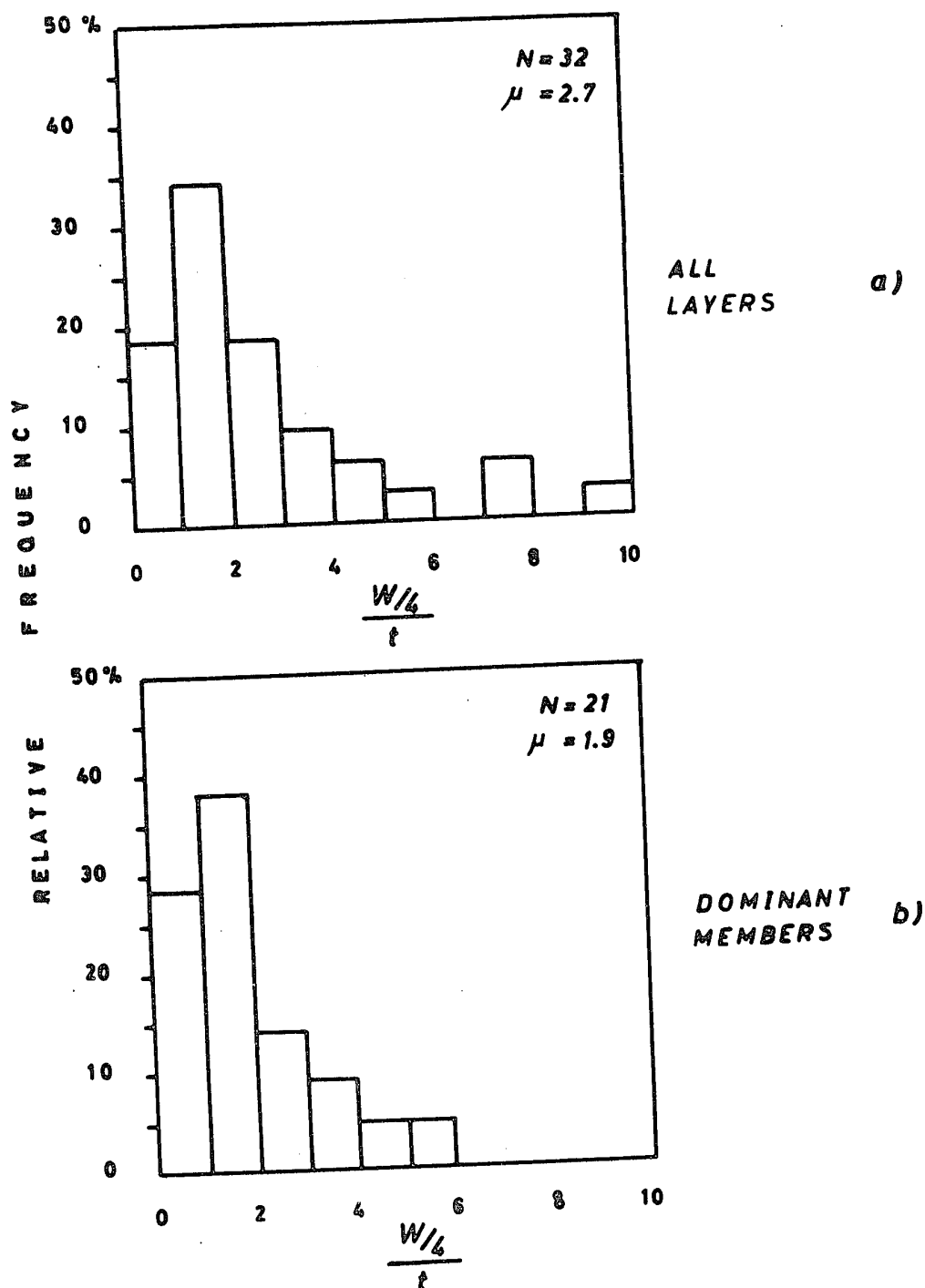


Fig. 17 Relative frequency of ratios of quarterwavelength ($W/4$), to layer thickness (t). N = number of layers measured. μ = mean value.

a) All 32 folded layers in 3 F_1 and 10 F_2 fold specimens of quartz or calcite - rich layers interbedded in pelite.

b) 21 dominant members in 3 F_1 and 10 F_2 fold specimens.

least squares method (see Appendix III) and its correlation coefficient determined (see also Appendix III). Its value (0.96) indicates a good fit of the data to a straight line.

For comparison the trend of the data from Currie et al. (1962) are shown in the same diagram. The ratio of the mean quarter-wavelength to mean layer thickness (1.58)* of the present work differs markedly from the ratio calculated by Currie and others ($W/t = 27$, which in terms of quarter-wavelength is $27/4 = 6.75$).

The discrepancy of results can in part be ascribed to the fact that the folds studied by Currie and others are mesoscopic and larger folds, and they could have a different W/t ratio from microscopic folds largely measured in the present work. A scale effect on the $(W/4)/t$ ratio can also be shown in the folds analyzed: the ratio $(W/4)/t$ for small folds (approximately 1) is smaller than for larger folds (almost 2). Mathematically this is due to the fact that the regression line has a rather greater intercept (see Appendix III).

A second more important factor to account for the difference in ratios is the relative small amplitude of the folds measured by Currie and others, whereas the folds studied in the present work have rather large amplitudes (when the amplitude of a fold increases considerably the value of W the

* The mean value 1.58 is given by the equation $\frac{\sum (W/4)_i / n}{\sum t_i / n}$ and it represents the ratio of the two means. On the other hand, the value 1.9 is given by the equation $\frac{\sum (W/4)_i}{n} / t_i$ representing the mean ratio. In general, these two quantities are not equal to one another.

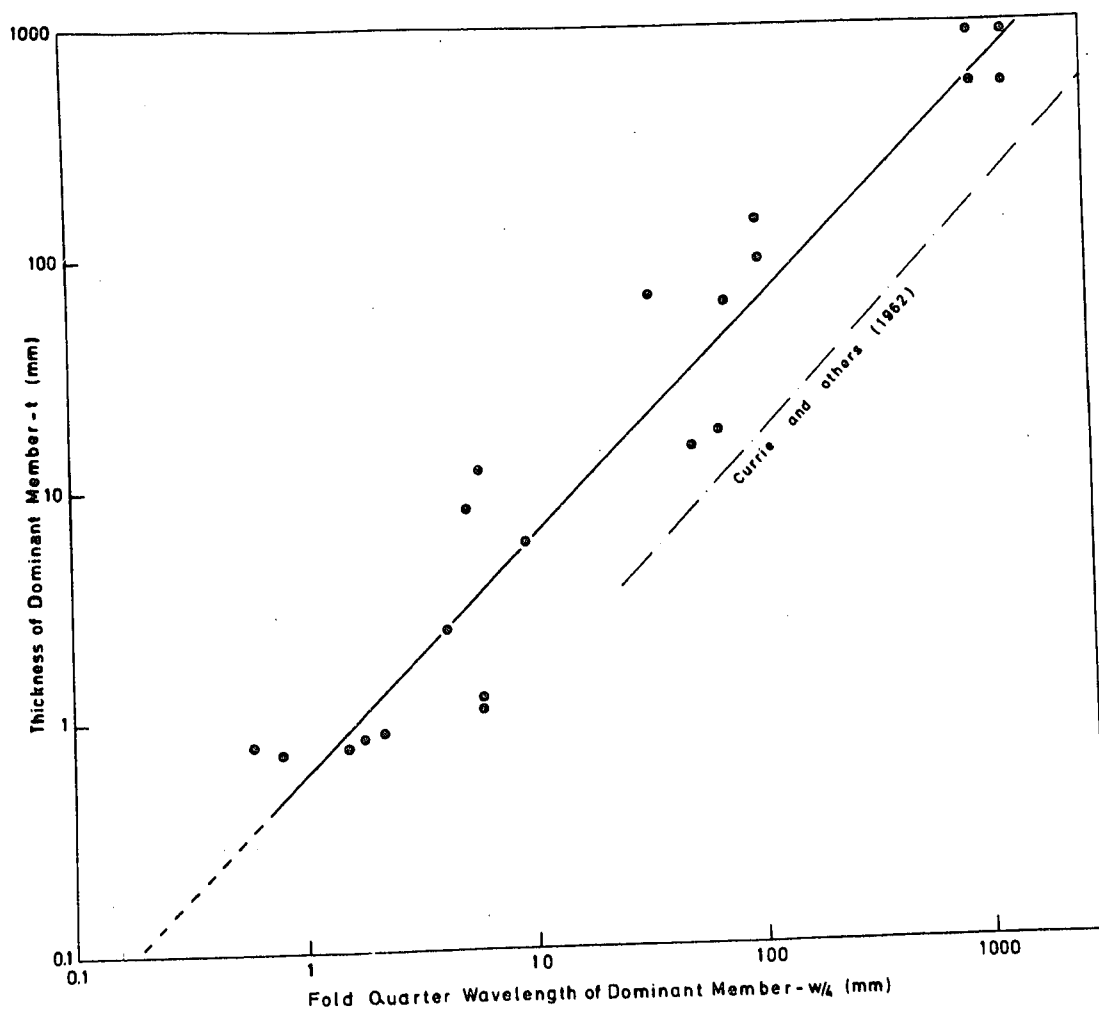


Fig. 18 Logarithmic relationship of fold quarterwavelength to thickness of 21 dominant members ($w/4/t \approx 1.58$) of the fold specimens measured in Fig. 17b.

final wavelength of the fold becomes smaller than W_d , the initial wave length generated more rapidly than the others at inception of buckling, and it can be shown that the ratio W/t is reduced up to $\frac{1}{2}$ of the ratio W_d/t for a total shortening of 50%, Currie et al., 1962).

A more consistent measurement, regardless of the amplitude, may be the arc length of the fold, which for very low initial amplitudes approaches W_d ; however, in the present work the wavelength was measured to make the results comparable with those of Currie et al. (1962).

A third possible explanation for the comparatively low values of the measured ratios of $(W/4)/t$ could be that the folds studied by Currie et al. (1962) were nearly true parallel folds, whereas the folds of the present investigation underwent a variable amount of flattening, which modified the fold shape, reducing the wavelength. If ideally the flattening component could be eliminated, the points in the diagram should shift parallel to the abscissa, giving a new trend closer to the line of Currie et al. (1962). Unfortunately the effect of flattening on $(W/4)/t$ ratio cannot be tested numerically because the flattening percentage in the folds holds only for the competent members and not for the whole folded sequence.

Viscosity Ratio and Wavelength-layer Thickness Ratio

From the Biot (1961) equation ($W_d = 2 \pi t \sqrt[3]{\mu_1/6\mu_2}$, see introduction p.70) it is possible to calculate the viscosity ratio of a folded single layer and of its surrounding medium, provided the ratio W_d/t is determined.

For the three folds studied in the present work which exactly satisfy Biot's equation assumptions (single layer embedded in a less

viscous medium, folding produced by buckling, true parallel folds), the fold arc length was calculated as an estimation of W_d , and the viscosity ratio calculated. In each case the value was not more than 25 (18, 23, 25).

These figures seem too low, because, according to Biot (1961) a viscosity contrast of 100 is necessary before buckling of a layered sequence can occur. With such a small viscosity ratio, deformation should be accomplished mainly by parallel layer shortening.

However, it is necessary to take into consideration that the viscosity ratio can be determined by the Biot equation only if the rock behaves as a viscous fluid. If the strain rate was high, rocks are unlikely to behave in such a way and the initial conditions necessary for the equation do not apply.

Flattening

For most of the competent layers forming class 1C folds the amount of flattening was calculated using the method of Ramsay (1962, 1967). In Table 5 the amount of flattening for each single layer is listed. This value, with few exceptions, represents the mean between the measurements of the two limbs considered separately. In Table 6, p.188 the average value of flattening for each folded sequence is given, and in Fig. 19b the relative frequencies of flattening for 71 layers are plotted using a class interval of 10. In all these tables and diagrams flattening is expressed, following the method of Ramsay (1962), in terms of percentage of shortening perpendicular to the axial surface of the folded layer (Fig. 15). This kind of strain measurement was preferred to the

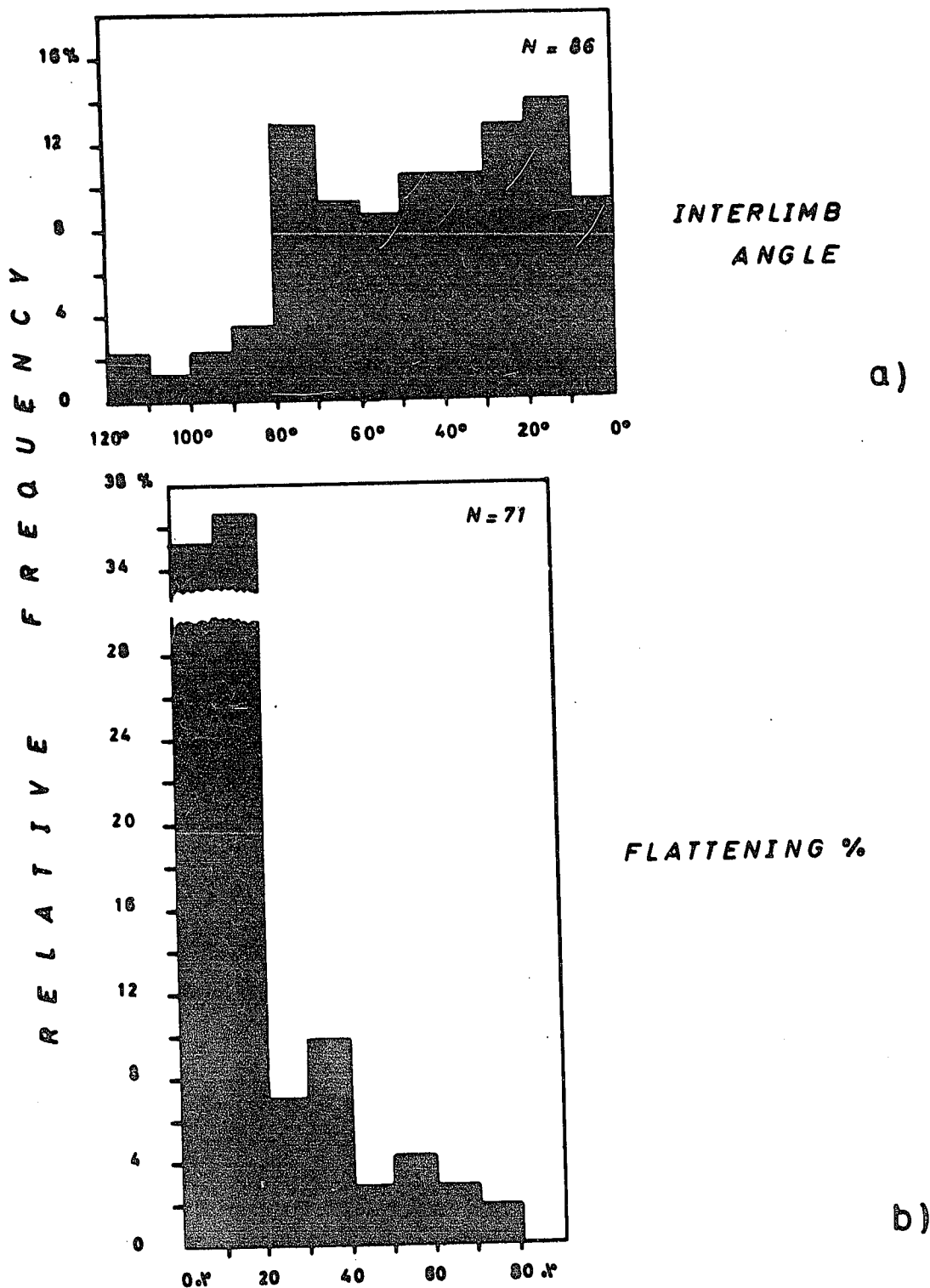


Fig. 19 Relative frequencies of interlimb angles and percentage of flattening of type 1C folds layers in 7 F_1 and 27 F_2 fold specimens. Mainly quartz or calcite-rich layers interbedded in pelite.

a) Interlimb angles grouped in classes of 10° .

b) Percentage of flattening grouped in classes of 10%.

computation of the quadratic elongations (Ramsay, 1967) because percentage of shortening can be easily visualized.

The frequency distribution of flattening is strongly skewed (Fig. 19b). The mean value for all the layers is almost 20%, and 70% of the layers underwent an amount of flattening between 0 and 20%.

In the graphs constructed in order to classify the folds and calculate the flattening for class 1C (Fig. 20) the lines joining the points with coordinates α and t' do not generally follow the smooth theoretical paths for the classes and subdivisions (Fig. 15). Also irregularities for one layer are not reflected in curves for adjoining layers in the same folded succession. These irregularities therefore appear to be due to original variations in thickness of the beds, as observed away from hinge areas of folds.

Because of such irregularities the measurement of flattening in many cases is approximate and an error of 25% is very probable.

It is of interest to note that the fold hinge is generally a point of elastic or viscous instability (Ramberg, 1964), which could coincide with points of original layer thinning. Such sedimentary thinning coinciding with fold hinges can be observed in several folded layers. In other folds the apparent thickening of the hinge areas may not take into account that these were originally thinner than the limbs. Thus, at least in some cases, the calculated amount of flattening represents a minimum.

The amounts of flattening calculated are not readily compared with fold flattening in other areas, as only a few determinations in rocks of similar type and metamorphic grade have been published, (for example Rust, 1965). A possible comparison is with measurements from high grade

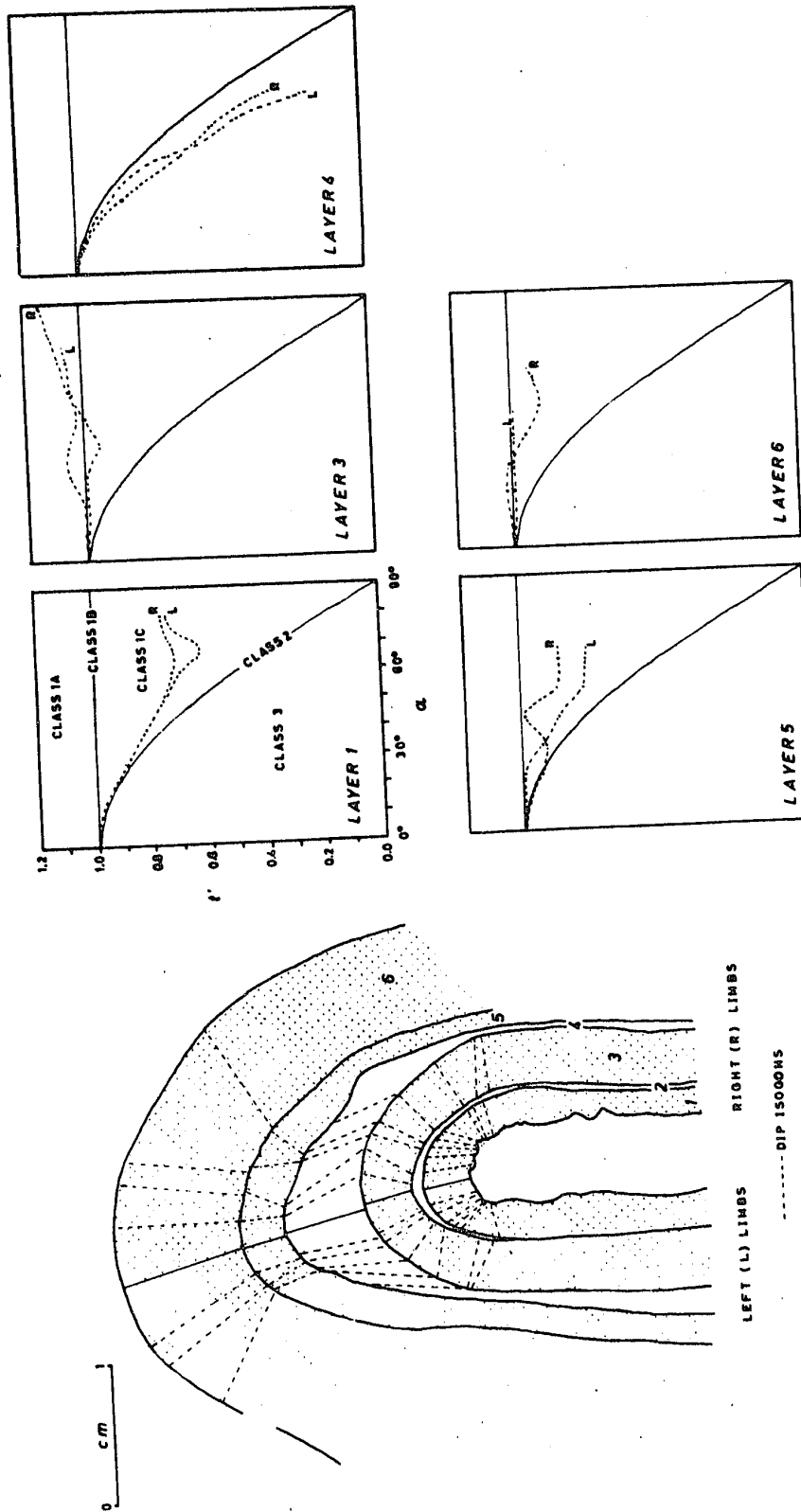


Fig. 20 F_2 fold profile and graph of t' (t_0/t) versus dip α (see Fig. 15) for 4 calcite-rich (dotted) layers and a pelite-rich layer. Lines joining the points with coordinates for t' and α do not follow the theoretical path of Fig. 15 due to original irregularities in thickness of the layers.

metamorphic terrains. For example Divi (1972) obtained from folded Precambrian Grenville rocks in the amphibolite facies a mean value of 50% flattening, which is more than twice the mean value of the present work. However the variations in flattening are wide, and further information is necessary for meaningful comparisons.

Interlimb Angle

In Table 5 the interlimb angles of 86 folded layers are listed and in Fig. 19a the same values are represented by a histogram, where the interlimb angle is grouped in classes of 10° . From the histogram the folds may be grouped as follows:

Fold Type	Interlimb Angle	Percentage
Isoclinal & Tight	$0^{\circ} - 30^{\circ}$	36%
Close	$30^{\circ} - 70^{\circ}$	39%
Open & Gentle	$70^{\circ} - 140^{\circ}$	25%

The interlimb angle varies widely throughout the area (details described later), and open and tight folds of the same generation occur in the same outcrop. Nevertheless, as shown above, 75% of the folds have an interlimb angle less than 70° and therefore are close to isoclinal. Most folds observed in the field are of this type.

Relationships between Flattening, Interlimb Angle and Other Parameters

The estimated flattening of the 1C folds could be related to the degree of folding as reflected by the interlimb angles.

However a graphical comparison does not reveal a linear correlation (Fig. 21). The graph shows that in folds exhibiting a relatively small

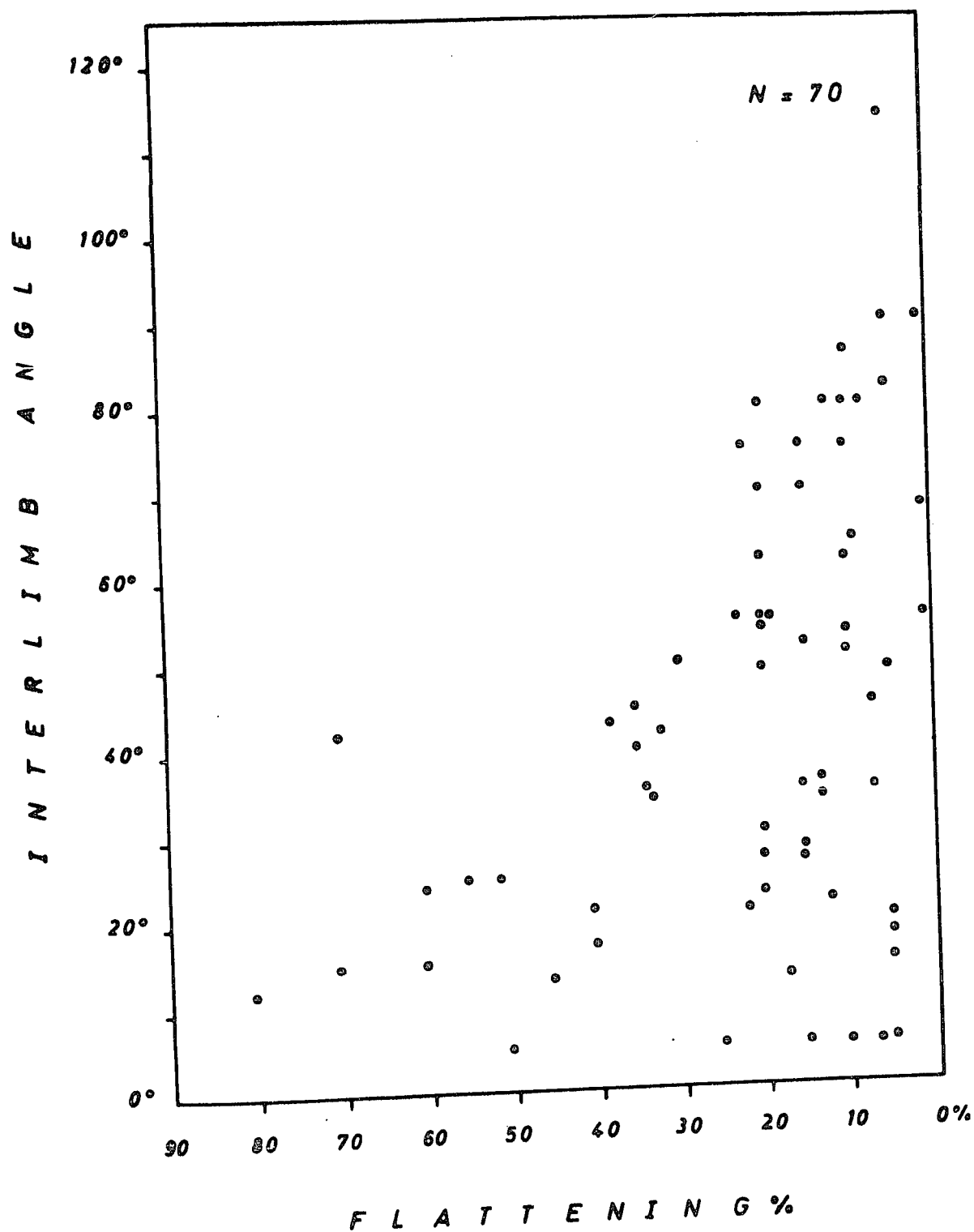


Fig. 21 Relationship between percentage of flattening and interlimb angle for 70 folded layers in 7 F_1 and 23 F_2 fold specimens. Lithology as in Fig. 19.

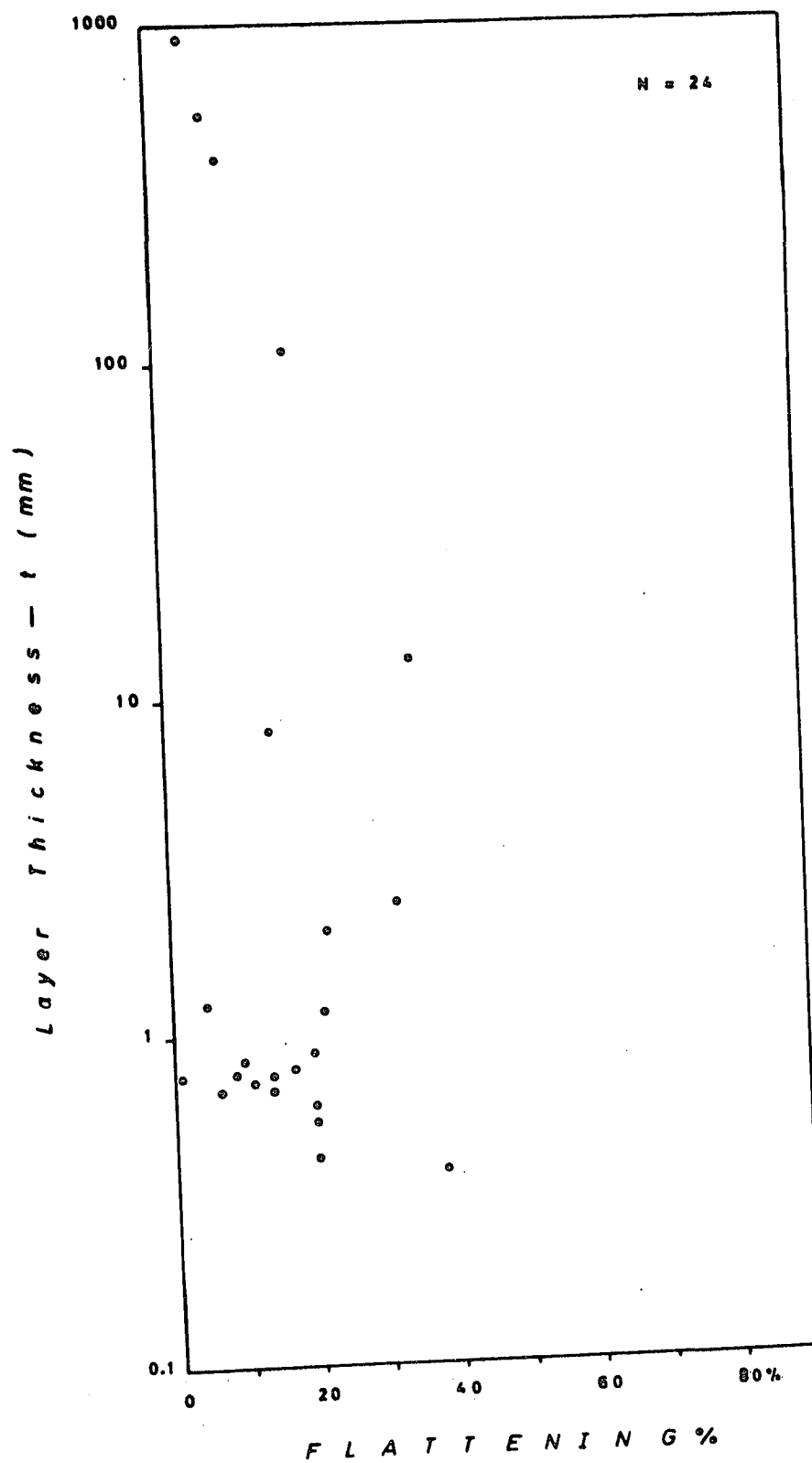


Fig. 22 Relationship between percentage of flattening and layer thickness for 24 folded layers in 3 F_1 and 10 F_2 fold specimens.

amount of flattening (0-30%) the interlimb angles vary widely from 5° to nearly 115° , whereas in folds with higher values of flattening (30-80%), the interlimb angle is usually less than 45° .

The wide variation in interlimb angles of folds with similar low percentages of flattening suggests that flattening started at various times during the buckling process, even when folds were still open.

Flattening was also compared with the thickness of the layer (Fig. 22), the wavelength and the amplitude, but no consistent trends were obtained (wavelength versus interlimb angle was also plotted without significant results).

In conclusion, the factors which control the relative amount of strain accomplished by buckling and by flattening, apart from that of ductility contrast of adjacent layers, are not readily apparent.

Comparison of Interlimb Angles and Flattening between F_1 and F_2 folds

It is of interest that workers elsewhere (Means, 1963, 1966, Park, 1969, Williams, 1970) have come to the conclusion that fold style is not an adequate criterion to separate fold generations. However, the definition of style* although supposedly all inclusive is usually restricted to

* The introduction of the term "tectonic style" has been attributed to Lugeon (Turner and Weiss, 1963, p. 79) and it embraces all the morphological features that characterize a given structure. For folds, for example, the shape, wavelength, amplitude, the presence or absence of lineations and foliations and their nature are the most common attributes of style. Additionally, Williams (1970) considers that metamorphism associated with the folds is part of the style.

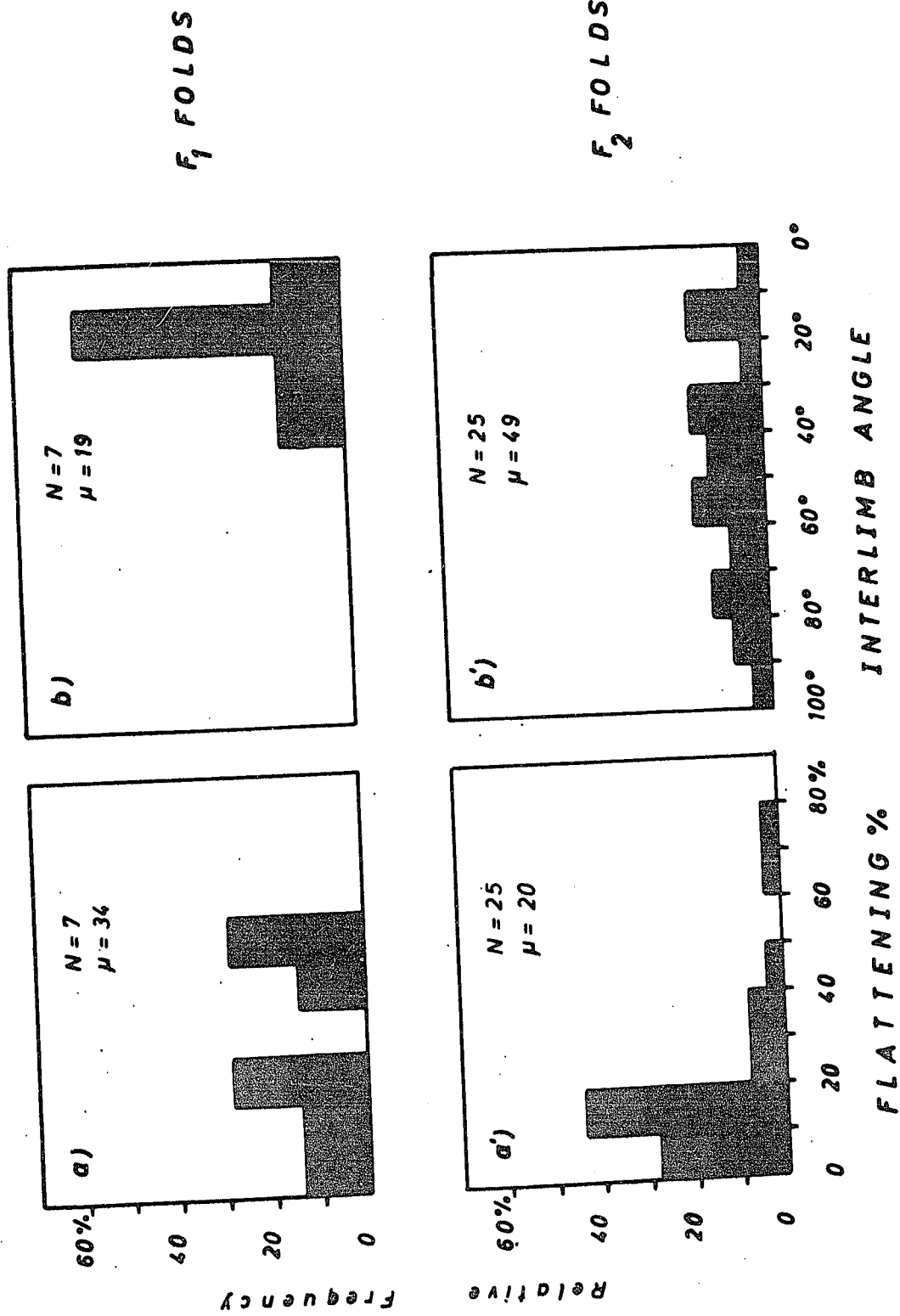


Fig. 23 Percentage of flattening and interlimb angles in F_1 and F_2 folds.
a, a') Relative frequency of flattening in 10% increments, 7 F_1 and 25 F_2 .
b, b') Relative frequency of interlimb angle in 10% increments, 7 F_1 and 25 F_2 .

general features of shape and size.

The interlimb angles and the calculated amount of flattening of F_1 and F_2 were compared in terms of relative frequency (Fig. 23) to see if these numerical parameters were statistically distinct for the two generations distinguished by other criteria (Chap. III).

The main limitation of this survey is the restricted number of data used, in particular only 7 F_1 folds were measured, each affecting a limited number of layers (2 or 3). For the more numerous F_2 folds (25) each value used is the average of several layers. If we take the mean value for each histogram (Fig. 23 a, a', b, b') we have:

	F_1	F_2
Flattening	34%	20%
Interlimb Angles	19°	49°

On average the F_1 folds are obviously more flattened and with a smaller interlimb angle than F_2 folds. As shown by the pairs of histograms, the overlapping of values is too great to use these fold parameters to separate the two fold generations with reasonable confidence. Nevertheless, in terms of the physical history of the rocks, the comparisons are useful.

Areal Variation of Interlimb Angles and Flattening in F_2 Folds

The areal variation of interlimb angles and amounts of flattening of the small-scale F_2 folds as measured in the laboratory and estimated in the field are shown in Figs. 24 and 25.

For the folds studied in laboratory the mean value of interlimb angle

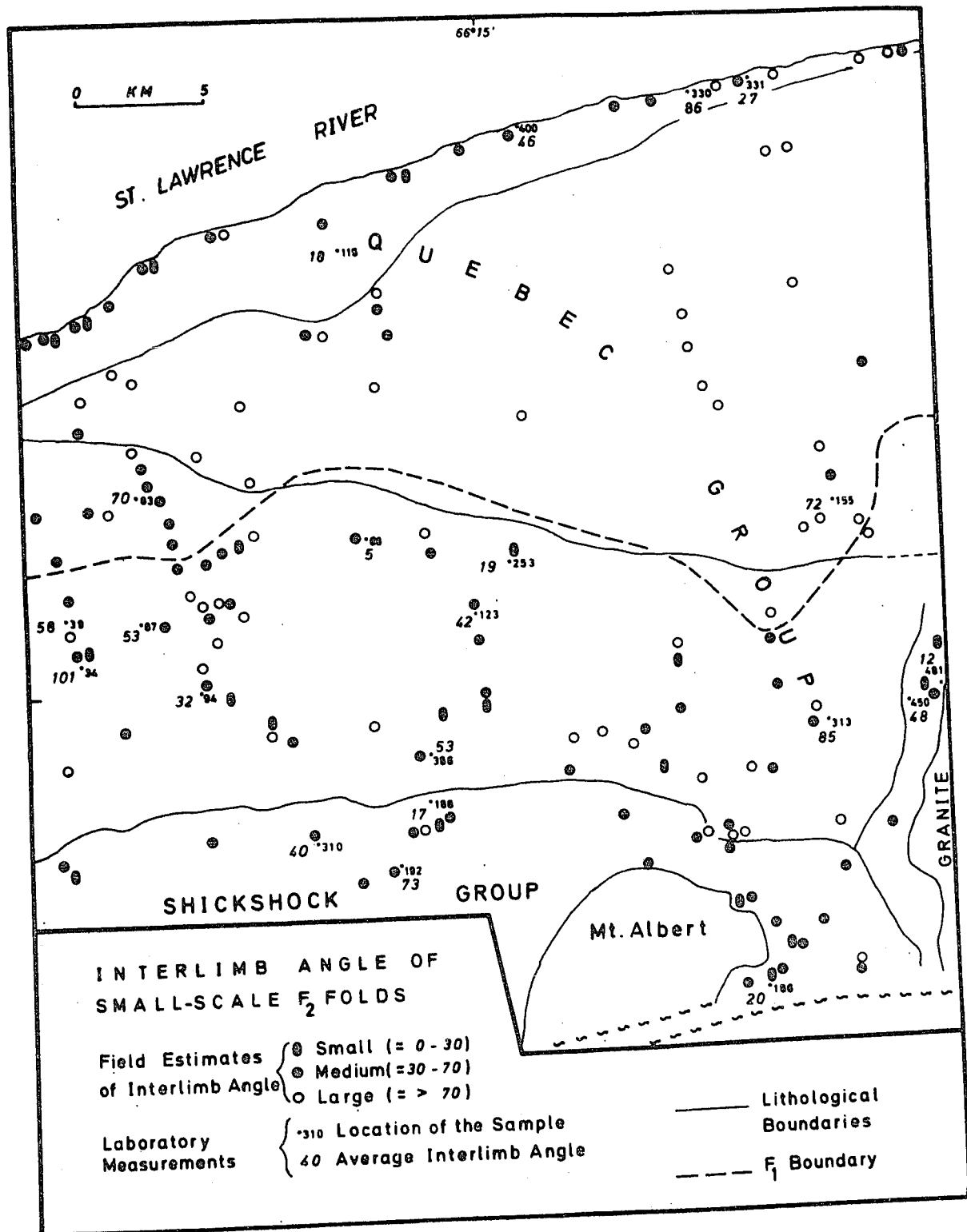


Fig. 24 Map showing the areal variation in interlimb angle of small-scale F_2 folds. Data from field estimates and laboratory measurements.

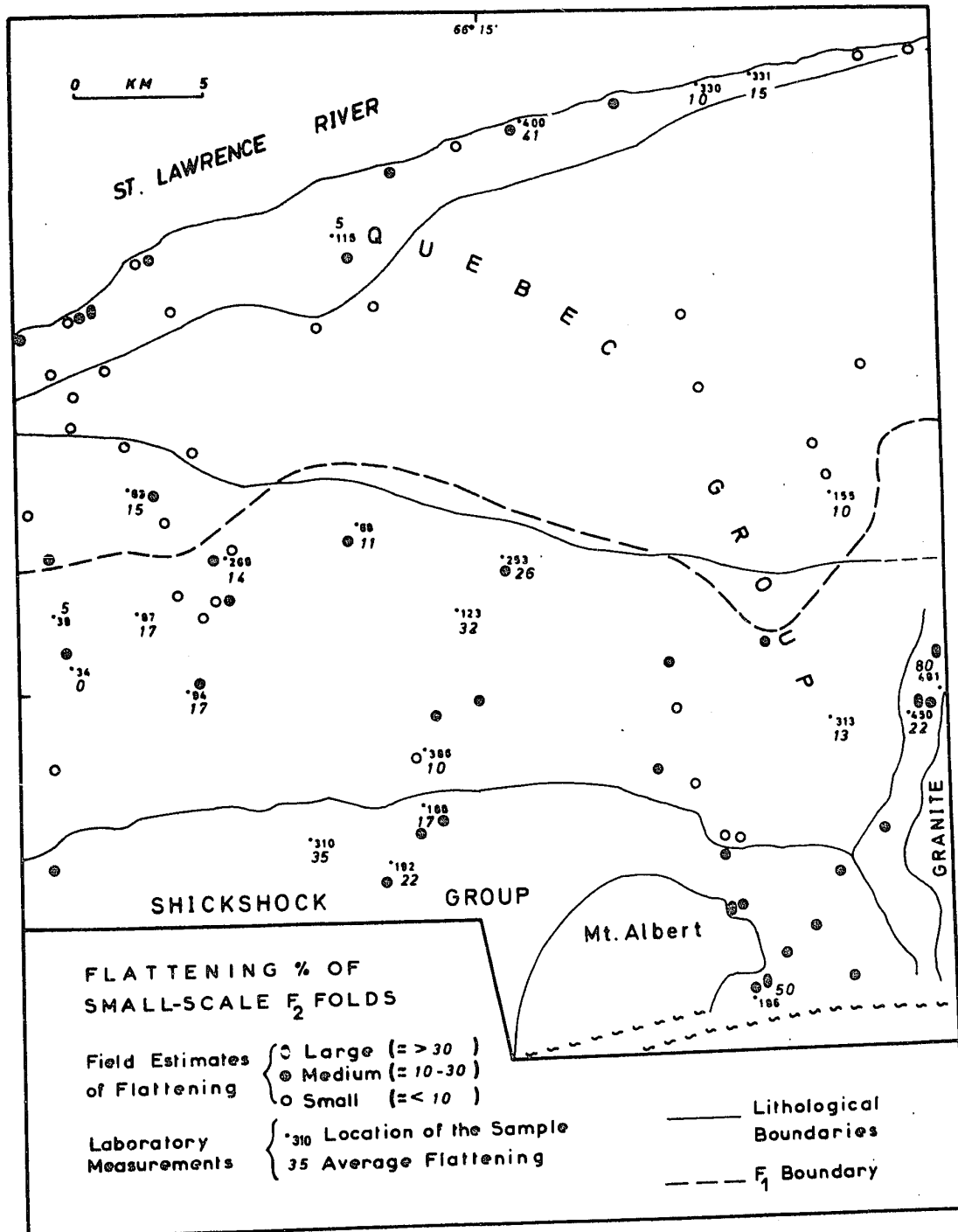


Fig. 25 Map showing the areal variation in percentage of flattening of small-scale F_2 folds. Data from field estimates and laboratory measurements.

of all the folds outcropping in the same locality are shown in the map of Fig. 24. For the folds observed in the field only semi-quantitative estimates were made. The interlimb angle was grouped in three classes: large (70° or greater), intermediate (about 30° to 70°), and small (less than 30°). Because folds frequently vary in tightness in the same outcrop or locality the approximate modal class of each locality was plotted on the map (Fig. 24).

The field estimation of flattening was more difficult; nevertheless, by comparison with the laboratory measurements of folds of similar shape, the folds were divided into three groups: folds with small (less than 10%), intermediate (about 10-30%), large (over 30%) amounts of flattening (Fig. 25). To the first group belong folds that do not show appreciable thickening of the hinge in comparison to the limbs. The second group is characterized by folds that clearly belong to class 1C. Folds which approach class 2 (similar) generally have an amount of flattening greater than 30% and are ascribed to the third group.

The interlimb angle varies greatly between adjacent localities (Fig. 24). Nevertheless folds in massive Middle Ordovician siltstones (map unit 3) are generally open, particularly in the eastern part of the study area. In the underlying Lower Ordovician sandstones, argillites etc. near the coast (map unit 2b) and in the phyllitic slates to the south (map unit 2a), tight and open folds are associated and do not show any systematic pattern. In the metavolcanics of the Shickshock Group open folds are rare, but the proportion of tight folds ($0-30^{\circ}$) is about the same as in units 2a and 2b.

The lithology appears to have controlled the interlimb angle in the

massive Middle Ordovician rocks and the presence of adjacent thick beds in unit 2a and 2b may account for relatively open folds in these rocks. In other respects, the local variation is greater than any apparent regional trend.

Flattening (Fig. 25) seems to have a spatial distribution fairly similar to that of the interlimb angle: low values throughout the Middle Ordovician siltstones and a large variation elsewhere. Flattening values lower than 10% were not observed in the metavolcanics of the Shickshock group possibly as a reflection of the lack of layers of high relative competency, such as sandstone in pelites. The two highest values of flattening measured are from near the Mt. Albert intrusive (50%) and near the Table Top granite (80%). It is tempting to suggest extra flattening near the Mt. Albert pluton during F_2 folding, as this is not incompatible with the textural evidence (Chap. IV). However, the Table Top granite is later than the F_2 folding, and the few values may not be representative.

Summary

The following summarizes the most significant results of the fold analysis:

- 1) In all terrains of the study area, in both F_1 and F_2 folds the most common types are class 1C in the competent and class 3 in the incompetent layers.
- 2) The dominant mechanism of folding appears to be buckling with an element of homogeneous strain (flattening). In the competent layers forming class 1C folds, the flattening is highly variable, but a majority of those measured

are not flattened more than 20%.

3) As observed in relatively low amplitude folds from other localities (Currie et al. 1962) wavelength and layer thickness of dominant members show a linear correlation. But the ratio $(W/4)/t$ (1.58) is smaller, partly because folds in the study area have rather large amplitude and partly because they are flattened.

4) The calculated flattening is not directly related to the interlimb angle, or to the wavelength and layer thickness.

5) On the basis of the means values, F_1 folds are tighter and more flattened than F_2 , but these two attributes are not sufficient to distinguish between the two fold generations.

6) Except for open small-scale F_2 folds that are moderately flattened in the synclinal area of Middle Ordovician massive siltstones, the local variations in interlimb angles and flattening percentage are greater than regional variations.

VI CLEAVAGE TYPES IN SEDIMENTARY AND LOW GRADE METAMORPHIC ROCKS

Introduction

This chapter deals with the optical analysis, classification, and interpretation of the various cleavage types observed in the sedimentary and very low grade metamorphic rocks of the Quebec Group. Cleavages in pelitic and semipelitic rocks are considered, but not those in limestones and quartzites where equidimensional grains predominate.

The nature and origin of cleavage has been a subject of considerable controversy and interest for over a century. However, field observations, laboratory experiments, and numerical methods carried out in the recent years (Cloos, 1947, Ramberg, 1963, Dieterich, 1969, Hobbs, 1971) suggest that most types of cleavage develop perpendicular to the local direction of maximum total shortening. Nevertheless, the internal processes that contribute to the formation of the various types of cleavages are not well understood.

Regardless of the relationships between stress, strain, and mechanism of cleavage development, most cleavage types in the study area fall into three well known groups defined (following Turner and Weiss, 1963, p. 98) as follows:

Slaty cleavage. This term denotes planar parting characteristic of slates and phyllites. It is independent of bedding and reflects a highly developed preferred orientation of grain boundaries, especially those of tabular crystals of micaceous minerals.

Fracture cleavage is a parting or incipient parting defined by closely spaced discrete parallel fractures, which ideally are independent of any planar preferred orientation of grain boundaries that exist in the rock.

Crenulation cleavage is a foliation defined by parallel narrow domains or discrete surfaces of incipient transposition of preexisting foliation.

However, some observed cleavages do not fit simply into the above definitions and there are various types of slaty, fracture, and crenulation cleavages which have intermediate and other characteristics.

The aims of the present work are to describe, in more detail than in the chapter on the structural phases, the morphological features of the various types of cleavage in the pelitic and semipelitic rocks, and to clarify to what extent cleavage "style" (as summarized in Fig. 28) is typical of specific fold phases.

The investigation* is based on the optical study of more than 150 thin sections coming mainly from the pelitic or semipelitic rocks of the phyllitic slate unit (map unit 2a), and from the immediately adjacent siltstone unit (map unit 3): only a few thin sections from the coast rocks (map unit 2b).

* The present work was carried out before the publication of the recent paper on cleavage by Williams (1972). This work, in many aspects similar, differs mainly in the fact that in all the rocks studied by Williams, metamorphic segregation was the main controlling factor of cleavage development and morphology. In the Mt. Albert area, conversely, segregation was operative only in the southern phyllitic slates.

were studied (the optical work was done mainly to trace the northern limit of D_1 deformation).

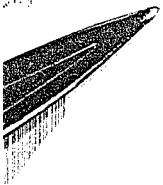
Slaty Cleavage Group

In many cases highly pelitic rocks of the phyllitic slate unit have an S_1 fabric defined by the preferred orientation of phyllosilicates (mainly chlorite and sericite), which is homogeneous throughout the investigated surface. The degree of orientation of the micaceous minerals was qualitatively estimated by optical methods (rotating the stage from extinction to maximum illumination), and, for a restricted number of rocks, by X-ray diffraction (see next chapter). This kind of cleavage is herein termed a slaty cleavage sensu stricto, (Plate 7a and 8a).

Where the rock contains appreciable amounts of quartz or calcite grain aggregates, chlorite and sericite wrap around the phenoclasts, and in the tails (pressure shadows) of larger grains they appear less oriented than elsewhere (Fig. 26 top, Plate 7b). Moreover, where the equidimensional grains represent 30% or more of the rock, the cleavage appears as a set of narrow foliae (Powell, 1969, Williams, 1972) or domains of less oriented phyllosilicates (formed partly by the merging together of the tails) that alternate with domains of better oriented platelets of micas. The surfaces bounding the foliae, frequently marked by iron stained material, are not planar, but as shown in Fig. 26, run irregularly, anastomosing, throughout the surface of a thin section. The thickness of the foliae varies from less than 0.1 to 0.5 mm, and is mainly dependent on the size of the equidimensional grains.

This kind of cleavage is partly discontinuous like fracture or crenulation cleavage. But as some alignment of grains is apparent throughout most of the rock,

Fig. 26 Cleavage types in sedimentary and low-grade semi-pelites
Drawings of thin sections. Top. Slaty cleavage sensu lato (S_2) defined by the alignment of sericite and chlorite in sandy shales of the siltstone unit (No S_1 present). Quartz grains (unshaded) with tails (pressure shadows) containing less well oriented flaky minerals. Tails merge in narrow domains or foliae bounded by iron stained surfaces that curve irregularly, anastomose and are discontinuous. Few pyrite (black) grains. (loc. 166).
Bottom. Narrow alternating domains of crenulation and slaty cleavage (S_2) crossing previous phyllitic cleavage (S_1) in pelites interlayered with calcite-rich beds (S_0). Platy minerals in S_2 slaty zones are oriented at small angles to the domain boundaries. Phyllitic slate unit, (loc. 290).



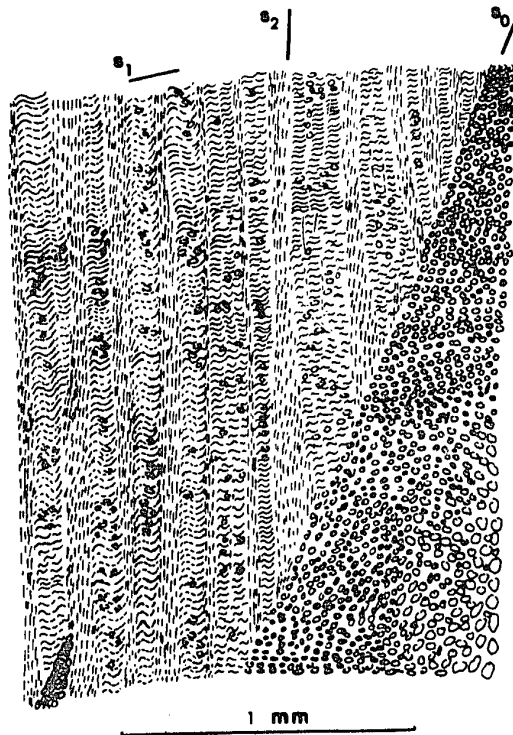
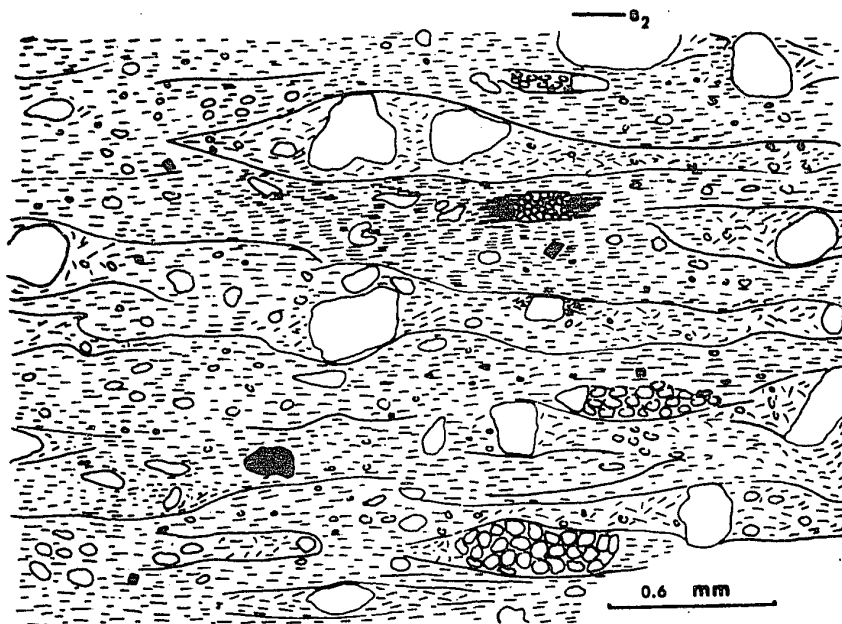


Fig. 26

it is here referred to as slaty cleavage sensu lato. A similar cleavage type has been described by Powell (1969) in shaly sandstones of the Siamo Slate unit (Michigan), and by Williams (1972) in the low-grade metamorphic rocks at Bermagui, Australia.

Another variety of slaty cleavage frequently observed in the phyllitic slates is characterized by closely spaced, discrete, parallel foliae which vary in composition to form laminations (Fig. 27a, Plates 8b, 9a). These result from the rhythmic alteration of calcite or quartz-rich layers or lenses, varying in thickness from 0.1 to more than 1 mm, with pelite-rich layers of approximately equal width. The phyllosilicates throughout the rock are strongly oriented parallel to the interfaces of the foliae, which locally are marked by iron oxides. In most cases, calcite and quartz grains are highly strained and lenticular; frequently, however, the grain size is so fine that detailed investigation is not possible.

This lamination type of cleavage generally appears to be S_1 (in 3-4 thin sections this lamination was observed parallel to S_2 cleavage; see for example Fig. 13c, p. 67). Because S_1 is in most cases parallel to the bedding, the lamination can easily be mistaken for an original sedimentary feature. However, a similar layering also occurs where the bedding is at an angle to S_1 (Fig. 27a, Plate 8b). Here lenses extend into the S_1 foliation from the oblique bedding, and grade into laminae.

As the phyllosilicates are well aligned throughout the rock, the cleavage is considered as a variety of slaty cleavage sensu stricto. Similar laminations have long been recognized in medium and high grade metamorphic terrains (gneissosity, metamorphic banding etc., Turner, 1941, Engels, 1959); more recently laminations oblique to bedding have

been described in almost unmetamorphosed slates (Talbot and Hobbs, 1968, Williams, 1972).

Fracture Cleavage Group

Relatively few examples of simple fracture cleavage were observed. More frequently, phyllosilicates are aligned along the cleavage surfaces which vary in spacing from approximately 0.1 mm to more than 1 mm. The proportion of aligned phyllosilicates varies from that where the planes are followed by a few flakes (Fig. 27b, Plate 9b) to (in rare examples) narrow bands of material displaying an original fabric, alternating with relatively wide domains of phyllosilicates aligned along the younger foliation, (Fig. 27c).

Fracture cleavage has been investigated by many workers (Leith, 1923, Billings, 1954, Baer, 1956, Fyson, 1962, Knill, 1961, de Sitter, 1964) and a rather confused terminology has been used (close-joints cleavage, false cleavage, fault-slip cleavage, etc.); however, more emphasis was given to the mechanical origin of this cleavage than to its morphological features. More detailed descriptions of fracture cleavage are by Crook (1964) and Beavis (1970); the "reticulate cleavage" of Crook and the "fissuring" of Beavis fall in the group of fracture cleavages of the present work.

Crenulation Cleavage Group

Crenulation cleavage also occurs in variable forms similar to that described elsewhere by several authors (Billings, 1954, Knill,

Fig. 27 Cleavage types in sedimentary and low-grade pelites and semi-pelites. Drawings of thin sections.

- a) Calcite (small circles) segregation laminae parallel to slaty cleavage sensu stricto (S_1) in pelites. Thin calcite-rich beds (S_0) are partially disrupted and transposed parallel to S_1 . Phyllitic slate unit (loc. 211).
- b) Fracture cleavage (S_2) slightly bent around larger (unshaded) grains in shaly siltstones of the siltstone unit. A few phyllosilicates flakes are aligned along the cleavage surfaces. (loc. 184).
- c) Uncommon type of "fracture cleavage" in pelites defined by narrow S_2 bands of aligned mica (mainly sericite) platlets crossing well-developed sedimentary parting S_0 defined by sericite and chlorite alignment. Siltstone Unit, sp. 173.

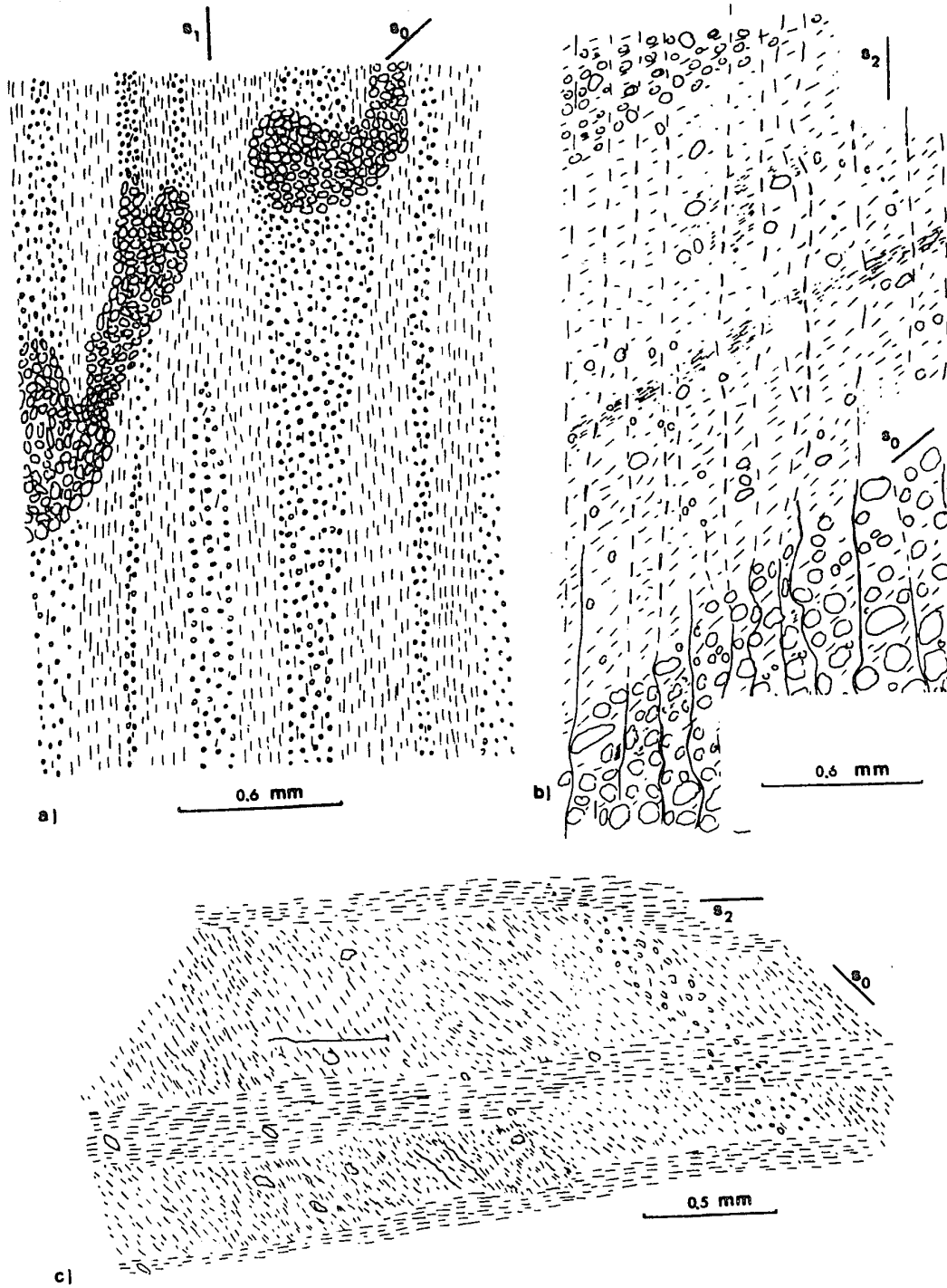


Fig. 27

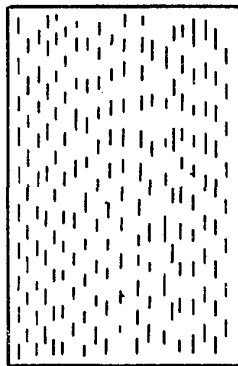
1960, Rickard, 1961, Turner and Weiss, 1963, Williams, 1972, etc.). Small crinkles of S_1 foliation grade throughout all the intermediate stages to disrupted microlithons separated by relatively wide domains of micaceous minerals aligned along or at small angles to the domain boundaries (Fig. 26 bottom). Segregation of calcite and quartz is also associated with this kind of cleavage (Fig. 13c, Plate 5b).

Cleavage Distribution

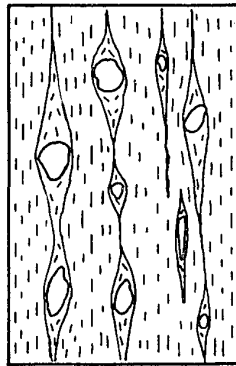
In Fig. 28, the various cleavage types observed in the sediments and low grade metamorphic rocks of the study area, are diagrammatically represented. It should be pointed out that not all these cleavages occur with the same frequency.

In highly pelitic phyllitic slates and slates of the Quebec Group (map-unit 2a), S_1 foliation is most commonly slaty cleavage sensu stricto (Fig. 28a). In such lithologies S_2 forms incipient crenulation cleavage (Fig. 28g), and also intermediate types between this and crenulation cleavage with slaty domains (Fig. 28h) the latter being uncommon. Slaty cleavage sensu stricto with segregation bands (Fig. 28c) was detected in not more than 6-7 thin sections. However, because in the phyllitic slates bedding is generally parallel to the S_1 cleavage, it was found difficult to separate a "metamorphic" from a sedimentary lamination parallel to the foliation. In semipelitic lithologies slaty cleavage sensu lato (Fig. 28b) and fracture cleavage with a few aligned phyllosilicates (Fig. 28e) are frequent. The cleavage formed by narrow bands of aligned silicates (Fig. 28d) is rare (2 thin sections), as is the fracture cleavage with no mineral alignment (Fig. 28f). The lack of a 'true' fract-

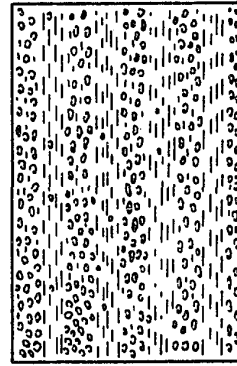
- Fig. 28 Schematic representation of the cleavage types recognised in the sedimentary and low grade pelitic or semi-pelitic rocks of the study area.
- a) Slaty cleavage sensu stricto (S_1), homogeneous mica alignment (Plate 7a, 8a).
 - b) Slaty cleavage sensu lato (S_1 and S_2), (Fig. 26 top, Plate 7b).
 - c) Slaty cleavage sensu stricto with calcite or quartz segregation bands (S_1) (Fig. 27a, Plate 8b, 9a).
 - d) Fracture cleavage defined by narrow bands of aligned mica platlets crossing earlier sedimentary parting (Fig. 27c).
 - e) Fracture cleavage surfaces along which a few phyllosilicates flakes are aligned (Fig. 27b, Plate 9b).
 - f) Fracture cleavage defined by spaced parallel fractures without mineral reorientation.
 - g) Incipient crenulation cleavage.
 - h) Crenulation cleavage defined by narrow alternating domains of crenulation and slaty cleavage S_2 crossing previous phyllitic cleavage (S_1) (Fig. 26 bottom).



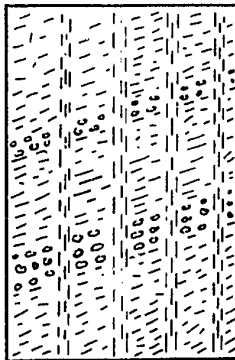
a



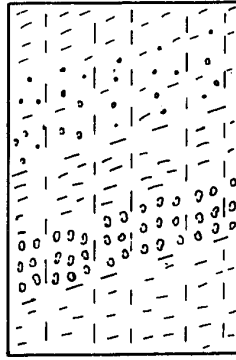
b



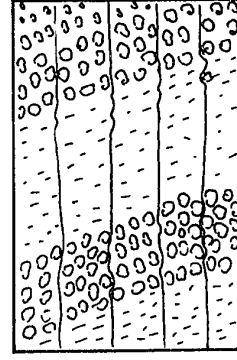
c



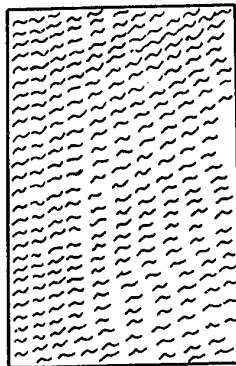
d



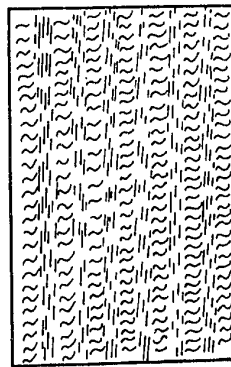
e



f



g



h

Fig. 28

ture cleavage is to be expected in the pelite-rich rocks studied.

Cleavage Type and Fold Generation

A simple relationship between cleavage type and fold generation is not apparent. Nevertheless, slaty cleavage sensu stricto, slaty cleavage with "metamorphic" segregation, and slaty cleavage, sensu lato, (Fig. 28a,b,c), are the typical S_1 foliation axial planar to the F_1 folds. On the other hand, fracture, crenulation, and slaty cleavage sensu lato form S_2 foliations axial planar to F_2 folds (Fig. 28d,e,f,g,h, and b). Thus, although slaty cleavage sensu lato occurs as both S_1 and S_2 , slaty cleavage sensu stricto is distinctive for S_1 .

Too few examples of cleavages were observed associated with F_3 and F_4 folds to warrant discussion beyond the fact that they are of the crenulation or fracture types. Thus, these types are not S_1 , but they could be S_2 or later surfaces.

Lastly, it is of interest to note that Williams (1972), carrying out a detailed study of the cleavage associated with two fold phases at Bermagui, Australia, did not find a simple relationship between cleavage "style" and fold generation either.

Discussion

The S_1 slaty cleavages are generally parallel to bedding on the limbs of F_1 isoclinal folds. Hence the cleavage is in rocks that appear to have undergone considerable strain. In contrast, the S_2 fracture and crenulation cleavages are more commonly at an appreciable angle to the

bedding (and where present, to S_1), and the F_2 folds indicate less deformation. Thus, the angular relations of the cleavages to bedding or other foliations, and the inferred strain, are considered as important factors in cleavage development. The actual mechanism of cleavage formation is more problematic and only a few aspects related to the development of slaty cleavage and segregation layering are considered here.

Several workers have recently described in detail and attempted to explain, the microscopic character of slaty-type cleavages in sedimentary and low-grade metamorphic rocks (Powell, 1969, Braddock, 1970, Clark, 1970 a, Williams, 1970, 1972). The role of mechanical reorientation of grains and crystals, and the effects of high fluid pressure (pore pressure) on the formation of the cleavage are stressed in support of the general theory of tectonic dewatering (Maxwell, 1962, Moench, 1966). Recent experimental work also indicates the importance of mechanical reorientation in producing aligned platy minerals (Means and Paterson, 1966). Additionally, Williams (1972) stressed the role of metamorphic segregation in the process of cleavage development.

Regarding the origin of the slaty cleavage with parallel segregation bands, it is useful to recall that Ramberg (1952) explained differentiation of mineral phases in terms of thermodynamic processes during diagenesis of sediments. According to this author, the total free energy for a certain mass of rock will decrease appreciably by the clustering of the different phases. Minute grains of calcite intermingled thoroughly with clay minerals will have greater molar free energy than an aggregate of calcite crystals in which calcite grains are surrounded by grains of similar species. It is well known indeed, that during

diagenesis concretionary differentiation occurs in many sediments (for example: calcite concretions in shales).

The hypothesis of the formation of slaty cleavage at an early stage of the deformation history, when the sediments were still unconsolidated (Maxwell, 1962), is not incompatible with the fact that "metamorphic" differentiation occurs parallel to the cleavage. Actually, the segregation of the more mobile minerals could take place in a late stage of cleavage development when the chemical and thermodynamic conditions were comparable to those of an advanced stage of diagenesis.

In the present case, the rock was originally calcite or quartz-rich beds alternating with pelitic-rich members in which calcite and quartz were disseminated. Apparently, during deformation, relatively mobile quartz and calcite components went into solution and nucleation occurred along variably spaced planes or zones parallel to the cleavage surface (cf. Williams, 1972).

It has been suggested that such a differentiation is controlled by shearing (Schmidt, 1932, Turner, 1941). Minerals with great plasticity concentrate in layers of intense shear motion, whereas less ductile minerals tend to remain in the less sheared portion of the rocks. In the present case, although some segregation of calcite and quartz is associated with crenulation cleavage for which some local shear motion may be inferred, more frequently differentiation is related to slaty cleavage. As shown in Fig. 27a, the segregation laminae of calcite do not lie in zones where the oblique bedding has been displaced along the cleavage more than elsewhere. Thus, the theory of segregation in zones of relatively higher shear does not apply. This supports the conclusions of

Talbot and Hobbs (1968, p. 586) that "differentiation" cannot always be correlated with differential strain.

Although the general relationships between deformation, slaty cleavage and segregation laminae are evident, and although in the present case temperature can be excluded as a major controlling factor, the ultimate nature of the processes appears to be obscure and certainly needs further investigation.

The presence of fracture cleavages with variable proportions of platy minerals aligned along the fractures may suggest that by progressive reorientation of the minerals along a new direction, they could be transformed into a type of slaty cleavage. However, the scarcity of fractures with appreciable alignment may indicate that slaty cleavage rarely forms through a fracture cleavage stage.

VII ANALYSIS OF PREFERRED ORIENTATION OF CHLORITE BY MEANS OF X-RAY DIFFRACTION

Introduction

The orientation of phyllosilicates is not always susceptible to measurements by optical means because of the small size of individual particles in many sedimentary and low-grade metamorphic rocks. Moreover optical means tend to be subjective and qualitative, whereas by using X-ray techniques it is possible to approach the problem from a quantitative point of view.

The following is an amplification of a paper on preferred orientations in the Mt. Albert area (Carrara and Fyson, 1971).

Previous Work

For some time engineers, soil researchers and physicists have been applying X-ray techniques to the study of preferred orientations of crystals and minerals in soils, metals, and rocks. More recently geologists have started to use these techniques for fabric analysis in rocks.

X-ray examinations of preferred crystallographic orientations in aggregates were initiated by Waver (1924), and Dawson (1927), who were among the first to study orientation in deformed metals. Sander and Sachs (1930) were the first to apply X-ray photographic techniques to study preferred orientations in rocks. Kratky (1930), Fairbairn (1943), and Ho (1947) also used X-ray photographic film techniques to study the

rock fabric.

More recently Decker, et al. (1943), Schultz (1949), and Field and Marchant (1949) developed Geiger counter X-ray diffractometer techniques to study preferred orientations. The introduction of the diffractometer constituted a considerable improvement in quantitative measurements of crystal orientation, but a complete three-dimensional orientation of crystals required the analysis of several sections and absorption corrections were needed.

Higgs, et al. (1960), following the technique of Jetter and Borie (1953), analysed several samples of quartzite, marble and limestone with spherical specimens mounted on a modified Norelco X-ray diffractometer. Silverman and Bates (1960) applied a transmission X-ray technique to study thin slabs of shales. Meade (1961, 1966) developed a method, similar to that used by Kaarsberg (1959), which is suitable for the analysis of phyllosilicates in soils or rocks with a well-defined foliation.

Similar techniques were used by Martin (1966) to study the orientation of kaolinite flakes. Engelhardt and Gaida (1963) used X-ray reflection to find the relationship between compaction of montmorillonite and kaolinitic clayey muds and the preferred orientation of these crystals. They demonstrated in this way that the orientation is mainly a function of the salinity of the solution.

Quigley and Thompson (1966) using similar techniques studied the relationship between clay platelet parallelism and void ratio of anisotropically consolidated natural and remoulded samples of Leda clay.

During the same period, the texture goniometer was used in reflection and transmission in order to obtain a faster and more complete spatial

orientation of crystals. Baker et al. (1969), with a modified Norelco pole figure goniometer in reflection and transmission, studied preferred orientations of quartz in fine-grained specimens of experimentally deformed and recrystallized flint. Oertel (1970), in investigating the deformation of a slaty, lapillar tuff, estimated the preferred orientation of the basal plane of chlorite and muscovite parallel to the cleavage surface to be 8-9 times random. Schwerdtner et al. (1970) applied the textural goniometer in reflection to analyze the alignment of hornblende grains in naturally deformed rocks. Clark (1970b), following the technique of Meade (1961, 1966), studied the effects of compaction, pure shear, and simple shear on the orientation of kaolinite in a kaolinite-water mixture with a high porosity.

Present Work

In the present work the preferred orientation of chlorite in undeformed and naturally deformed sedimentary and low-grade metamorphic rocks is studied by applying an X-ray method. The numerical results are then of use in distinguishing between parallel sedimentary (S_0) and tectonic (S_1) foliations.

With some modifications, the X-ray technique followed by Meade (1961) was chosen because:

- a) In the micaceous minerals the crystallographic and morphologic orientations coincide.
- b) In the specimens analyzed, the plane of maximum preferred orientation of the basal plane of the phyllosilicates was always macroscopically recognizable as a fissility plane or bedding plane. For this reason it was possible to analyze three mutually perpendicular sections (one of them

parallel to the foliation) as required by the technique and thus obtain a quantitative measurement of the preferred orientation of 001 plane parallel the foliation of tectonic or sedimentary origin.

c) The method requires a standard X-ray diffraction goniometer which was more readily available than a texture goniometer, and the sample preparation does not take too long.

Among the phyllosilicates present in the rock, chlorite was used because it is the most common phyllosilicate of the sedimentary and low-grade metamorphic rocks throughout the area, and frequently it is the most abundant in individual samples.

Experimental Procedure

Three mutually perpendicular sections are cut from the specimens. One of these is parallel to the bedding or cleavage plane and the other two are perpendicular to it and mutually perpendicular. The size of each section is approximately 30 by 40 mm. and the thickness is about 3 mm. (this is much greater than the maximum penetration power of the X-ray beam (0.5 - 0.8 mm)).

The surface of the section is well polished so as to obtain sharper peaks on the diffractometer chart and to avoid peak shifting. The polished section is then placed in a rotating specimen holder, which is mounted on a wide-range goniometer of the Philips X-ray diffractometer and exposed to iron filtered cobalt radiation. The sample is run from $13^{\circ} 2\theta$ to $16^{\circ} 2\theta$ and from $39^{\circ} 2\theta$ to $42^{\circ} 2\theta$, that is from a low to a high-angle reflection.

The running velocity is very low, $1/4^{\circ}$ per min. A xenon-filled

proportional counter is employed which is linear in response within a 4% error for the interval used.

For each chart the integrated area of the peaks of 002 and 131 for chlorite is calculated, and following Meade (1961), the ratio between the 002 and 131 peaks is computed for each of the three mutually perpendicular sections of each specimen. The ratio for the section parallel to the bedding or cleavage is then divided by an average ratio for the two perpendicular sections. Then, following Clark (1970b), the square root of the quotient is taken as a measure of the orientation parallel to the bedding or cleavage of the basal planes of chlorite, and it is called the preferred orientation index (POI).

If 002//, 131//, 002⊥A, 131⊥A, 002⊥B, 131⊥B, represent the peak intensities in the sections of the specimens, which are parallel and perpendicular (A,B) to the bedding or cleavage, then the preferred orientation index will be the following:

$$POI = \sqrt{\frac{002//}{131//} \cdot \frac{002\perp A + 002\perp B}{131\perp A + 131\perp B}}$$

From the above formula it follows that the POI = 1 when the chlorite flakes have a random orientation, and the POI > 1 when the chlorite basal planes have a preferred orientation parallel to the bedding or cleavage. Theoretically the POI can reach ∞ for a perfectly oriented aggregate of chlorite crystals.

The above "double ratio" has been selected because the factors (apart from the state of preferred orientation of the crystals) that affect the intensity of X-ray reflection, like density of crystals per unit area, particle size, chemical composition, and degree of crystallinity are cancelled out of the index.

The double ratio is particularly useful in the analyzed rocks because the composition of chlorite changes, even if only slightly, from specimen to specimen and the ratio between 002/131 intensities changes with chlorite composition.

The peak intensities of 002 and 131 (or 202 as in chlorite 131 and 202 planes give the same reflection) were chosen to measure the preferred orientation of chlorite for the following reasons:

- a) The 002 reflection in the chlorites of the analyzed specimens is the strongest basal plane reflection.
- b) The 131 (or 202) reflection has been used instead of the peaks of the planes which are cozoal with the c axis $hk0$, $h00$, $0k0$ because the reflections from these planes are too weak to be measured with sufficient accuracy. The angles in chlorite between the basal plane and 131 and 202 are approximately 80° and 70° , and, for the purpose of the present work, they are useful as much as the planes cozoal with the c axis.

Limitations of Method and Sources of Error

When applied to natural rocks, as in the present work, this technique involves a series of limitations and errors which should be taken into consideration.

- a) Samples which contain very small amounts of chlorite give variable

results because, due to the interference of the background noise, the measurement of the small peaks is difficult. Thus specimens with a low chlorite content can be compared with the specimens of high chlorite content in only a semi-quantitative way.

b) The same kind of inaccuracy in peak measurement arises when a specimen has a very high preferred orientation, because, in this case, the peak areas of 131 and 002 in the sections respectively parallel and perpendicular to the foliation or bedding are very small, even if the rock is very rich in chlorite (theoretically, the 002 peak should disappear for a perfect mineral orientation).

c) When used in aggregates of different mineralogical species the techniques can be affected by the overlapping and interfering of different peaks. In the specimens used for this work, no minerals are present which interfere with 002, but the intensity of 131 in some specimens, which are particularly rich in muscovite, is affected by 202 of muscovite.

d) Lastly, if the rock contains two or more planes of preferred orientation that are at a small angle (bedding, different cleavages), the peak intensities parallel and perpendicular to one plane are influenced by the particles oriented along the other planes, giving a POI which may be greater or smaller than the true value.

The following steps were taken to minimize the sources of error:

- a) Specimens with small amounts of chlorite have been disregarded.
- b) The measurement of small peaks was aided by changing the amplification factor during scanning of a section.
- c) Samples with a large proportion of muscovite were not used.
- d) Rocks with two or more foliations at small angles were not used.

Unfortunately, the cited limitations (particularly the last one) determined "blank spots" in the map showing the areal distribution of the measured chlorite fabrics (Fig. 29).

As a test of fabric homogeneity and analytical errors, for most samples 2 or 3 X-ray scans were made on different parts of a single section (after the first run, the sections were moved so that 60% of new surface was struck by the beam during the second scan). The resultant standard deviations (Table 3) vary in percentage from 0.7% to 19% with an average of about 5%. This local variation is considered acceptable for the purpose of the present work because the variation in POI among the various samples considered is of much greater magnitude. Thus these factors do not affect the general conclusions.

Data Treatment

Sixty samples of pelitic rocks were analyzed, (including 50 from the Quebec Group and one from a lens of phyllitic slates with the Shickshock Group), each with one of the three mutually perpendicular sections parallel to bedding (S_0), S_1 foliation or S_2 foliation (Table 3)*. It is apparent from

* More than 100 samples had originally been analyzed by means of the Philips X-ray diffractometer of the Geology Department of the University of Ottawa. Due to the faulty operation of that instrument a large non-systematic error was introduced in the results. Because of these instrumental limitations, 60 samples were rerun at the laboratories of the Soil Research of Canada Agriculture Department, and only these values were taken into account.

Table 3 Preferred Orientation Index (POI) for Chlorite in Cambro-Ordovician Rocks (Quebec Group), Silurian Sediments (St. Leon Formation) of Mt. Albert Area, and in Ordovician Rocks (Ottawa Formation) of Ottawa Area.

A Quebec Group POI parallel to bedding S_0				B Quebec Group POI parallel to bedding S_0 and tectonic foliation S_1			
Sample*	POI	σ	Lithology	Sample*	POI	σ	Lithology
1	1.02	0.01	calcareous shale	26	13.40	1.20	calcareous slate
2	2.70	0.36	calcareous shale	27	15.76	1.99	calcareous slate
3	2.02	0.23	calcareous shale	28	7.57	0.59	slate
4	1.30	—	calcareous shale	29	16.03	0.67	slate
5	1.35	0.97	shale	30	19.21	0.42	phyllitic slate
6	4.55	0.86	shale	31	4.50	0.04	slaty limestone
7	1.79	0.06	calcareous shale	32	14.98	0.48	slate
8	1.09	0.06	shaly limestone	33	22.73	0.22	phyllitic slate
9	5.83	0.42	shale	34	3.51	0.13	silty slate
10	4.82	0.23	shale	35	12.68	0.44	slate
11	5.01	0.02	shale	36	8.33	0.48	slate
12	2.14	0.06	silty shale	37	16.55	1.21	slate
13	1.41	0.02	silty shale	38	8.15	0.21	slate
14	2.68	—	silty shale	39	7.89	0.36	slate
15	0.70	0.01	shaly siltstone	40	10.24	1.05	slate
16	3.39	0.01	shale	41	4.42	0.37	silty slate
17	1.59	0.07	silty shale	42	10.91	0.23	calcareous slate
18	3.67	0.05	shale	43	10.73	1.03	slate
19	2.93	0.24	shaly siltstone	44	33.34	2.73	phyllitic slate
20	1.74	0.08	shaly siltstone	45	7.72	0.31	phyllitic slate
21	8.93	0.10	shale				
22	4.14	0.37	shale				
23	1.33	0.02	shaly siltstone				
24	2.01	0.05	shaly siltstone				
25	6.69	0.37	calcareous slate				

C St. Leon and Ottawa Formations POI parallel to bedding			D Quebec Group POI parallel to S_2 cleavage		
POI	σ	Lithology	POI	σ	Lithology
2.34	—	silty shale	4.25	0.07	shaly limestone
2.19	—	silty shale	2.41	0.06	shaly siltstone
3.50	—	shale	1.56	0.05	silty shale
5.47	—	shale	4.05	0.33	shaly siltstone
2.71	—	silty shale	1.11	0.01	shaly siltstone
1.55	—	shale	2.77	0.50	shaly siltstone
3.35	—	shale	3.44	0.05	calcareous slate

*Samples numbered as on map (Fig. 1).
 σ Standard deviation where listed, based on 2 or 3 scans of a sample.

the table that the more pelitic the rock is the higher is the POI both for sedimentary and tectonic fabrics. Nevertheless, the effects of lithology have been minimized by omitting non-pelitic samples.

Although in several outcrops it is not clear whether or not the bedding-plane tectonic foliation S_1 is present, field relations (Chap. III) and examination of thin sections under the microscope enable the study area to be separated into a southern portion in which S_1 is apparent in slates, and a northern portion where S_1 is absent in shales and argillaceous siltstone (Fig. 29). Most of the argillaceous rocks thus fall into two fabric groups that are geographically distinct: the southern tectonites with S_1 and parallel bedding S_0 , and the northern sedimentary rocks with only S_0 . A subgrouping of the tectonites into slates, and phyllitic slates farther south with a slightly larger crystals along S_1 is also possible, although such a subdivision is rather subjective (Fig. 29).

On the basis of geographical location and of numerical differences (see below) 45 POI measurements parallel to the bedding S_0 were divided into two groups: the first located in the northern sedimentary rocks S_0 (25 samples, Table 3A) and the second group (20 samples, Table 3B) from the tectonites where S_1 parallels S_0 .

These "sedimentary" and "tectonic" populations of POI values are also represented on histograms (Fig. 30 a,b). The mean of the POI values parallel to bedding S_0 is equal to 2.99 with a variance equal to 4.20 (Fig. 30b), and the mean and variance of the POI values parallel to S_0 and S_1 are 12.43 and 50.77 (Fig. 30a).

Although the means of these two populations differ, due to the relatively small number of data it is necessary to test whether or not the diff-

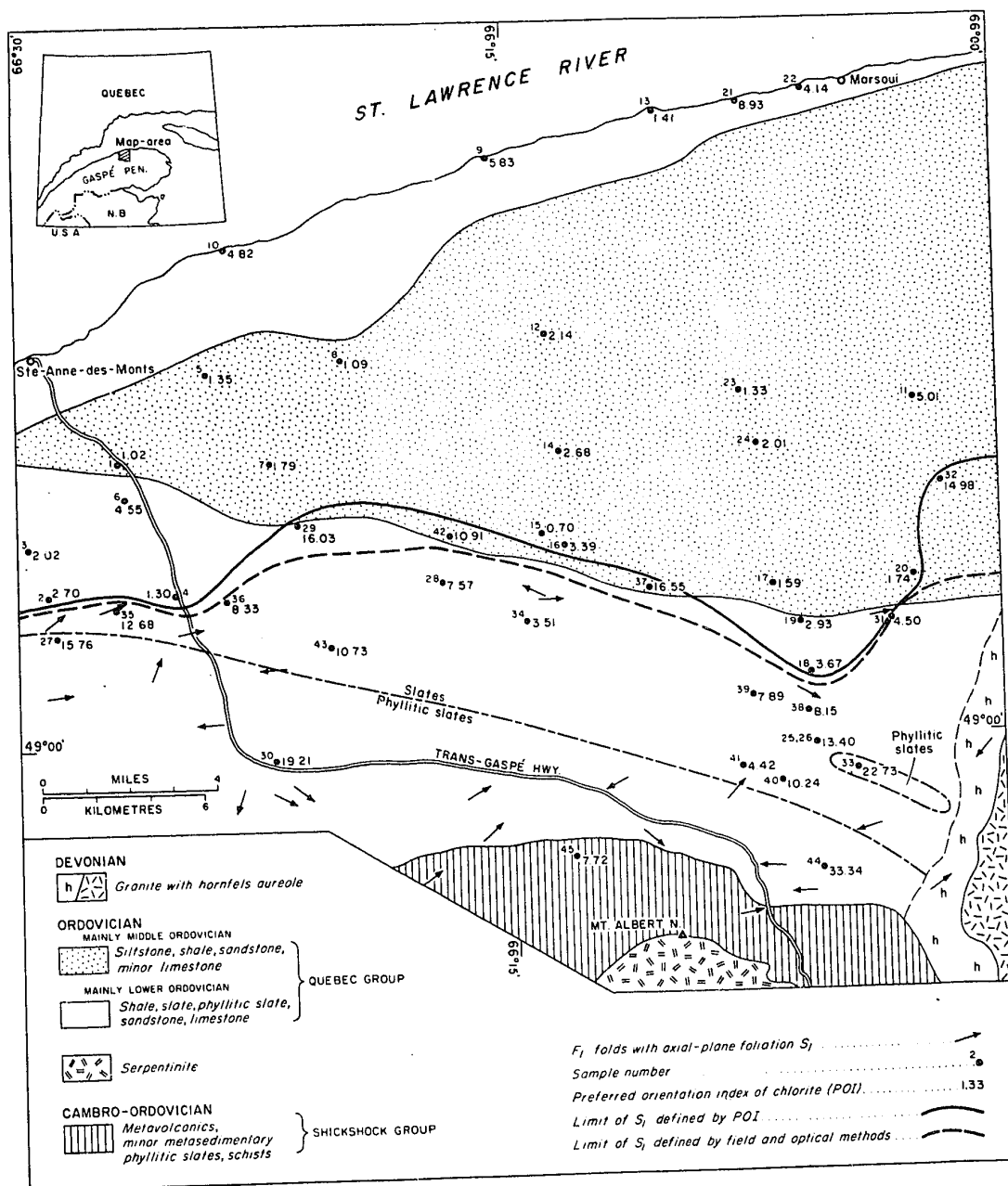


Fig. 29 Map of the Mt. Albert area with preferred orientation indices (POI) of chlorite.

erence is statistically significant.

The frequency distributions in both cases are strongly skewed and with variances that differ widely. Thus, it is best not to apply Student's t test directly. The POI values were therefore converted to their natural logs (Fig. 30 a',b'). In this way the population of POI parallel to S_0 and S_1 is much closer to a normal distribution, and the POI values parallel to S_0 are slightly closer, as proved by using the χ^2 test. Moreover the two log-populations have variances (0.323 and 0.429) sufficiently similar, as revealed by the F test, to apply with confidence Student's t test. Using the t test to compare means of the log populations, it is found that $t = 7.99$; the t distribution is 2.4 for 99% confidence level and the appropriate 43 degrees of freedom. Thus the hypothesis that the two means come from the same population is rejected, and on the basis of POI values the bedding foliation and tectonic foliation populations are statistically distinct.

From the histograms (Fig. 30 a,b) it is possible to select from where the two populations appear to overlap a POI value (about 7) that may be used as a boundary value. The boundary is such that if it is applied, there is minimum overlap. After initial grouping, this value was also taken into consideration in the placing of doubtful samples into the two populations, in particular 4 samples from near the north limit of S_1 , where by field and optical methods it was not readily apparent whether or not S_1 is present.

The boundary POI was then used in conjunction with field evidence, such as the presence or absence of F_1 hinges, to establish the northern limit of S_1 . As the selected value is an arbitrary cut-off between

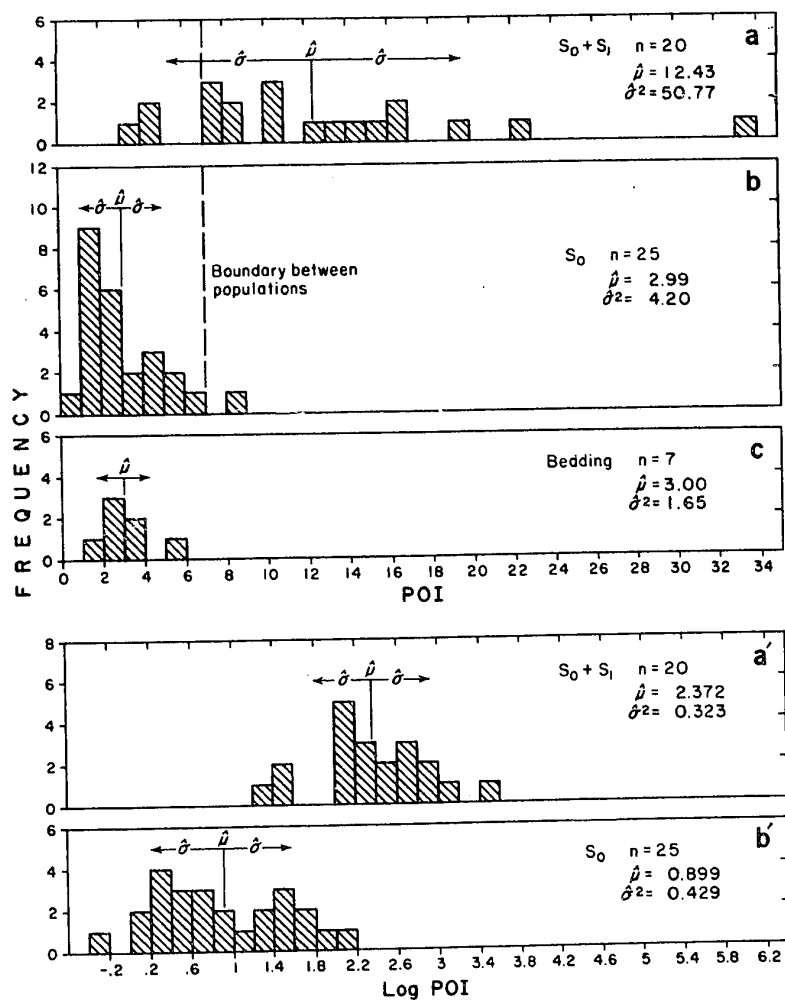


Fig. 30 Histograms of preferred orientation index (POI) for chlorite.
a) Parallel to bedding S₀ and tectonic foliation S₁. Quebec Grp.
b) Parallel to bedding S₀, Quebec Group.
c) Parallel to bedding, St. Leon and Ottawa Formations.
a', b' Natural logarithmic values for a and b.
μ̂ = mean
σ̂² = standard deviation

populations that slightly overlap, the line drawn on the map (Fig. 29) does not rigorously exclude all higher values from the low side and vice versa.

It is of interest to see whether or not the POI values have been affected by D_2 deformation, which produced the F_2 folds that are penetrative throughout the area underlain by the Quebec and Shickshock Groups. For comparison the values of POI parallel to bedding were measured from 2 samples of Silurian silty shales (St. Leon Formation) south of the Shickshock Fault, and from 5 samples of shaly rocks of the Ottawa Formation, near Ottawa, Ontario (Fig. 30 c). In both areas the beds dip at low angles and are virtually undeformed.

The mean POI (3.00) of the 7 samples does not appreciably differ from that of S_0 (2.99) from the folded rocks of the Quebec Group. Thus the POI for S_0 does not appear to be affected by the F_2 folding, and the degree of orientation may be comparable in silty and shaly rocks in other areas.

Discussion

Although the technique outlined has used POI values to distinguish bedding-plane foliations, it must be stressed that X-ray measurements without careful field and microscopic observations can be misleading. For example in some shaly siltstones of the Quebec Group (map unit 3), bedding is weakly defined, S_1 is missing and S_2 cleavage across the bedding is the most prominent foliation. Generally this cleavage is distinct from S_1 (Chap. VI), however, where bedding is obscure and S_2 is macroscopically penetrative, S_2 is not readily distinguishable from bedding-plane S_1 .

cleavage as developed in the southern slates. The POI values of chlorite parallel to S_2 in 7 samples analyzed vary from 1.11 to 4.25 (Table 3D), that is they are lower than the POI for S_1 ; but without field and microscopic observations the POI might easily be considered as indicative of an original fissility, S_0 .

The low POI values for S_2 cleavage is readily explained by the microscopic observation that in the samples analyzed it is a type of fracture cleavage or slaty cleavage sensu lato with relatively little orientation of the flaky minerals parallel to the cleavage surface.

It is of interest to determine the POI of chlorite parallel to S_1 where S_0 is not parallel. Unfortunately it is difficult because, due to the isoclinal nature of the F_1 folds, (see Chap. III p.28), S_1 is oblique to S_0 only in the hinge areas, few of which are exposed. However, in one sample of calcareous slate from near an F_1 hinge (Fig. 31a), the POI of S_0 (6.69) and oblique S_1 (4.16) were measured separately. Two meters away from the hinge, S_1 and S_0 are parallel (Fig. 31b) and the combined POI is 13.4. This is of the same order of magnitude as the sum (10.85) of the independently determined POI for S_0 and S_1 . If there is a simple addition to the POI for S_0 of an S_1 value that is approximately constant throughout, the presence of S_0 parallel to S_1 has not greatly enhanced the degree of chlorite orientation that can be attributed to tectonism.

It is obvious that this hypothesis, with its implications regarding the formation of foliation during folding, will have to be tested with additional measurements.

In nearly all the cases of shales with a bedding-plane fissility of

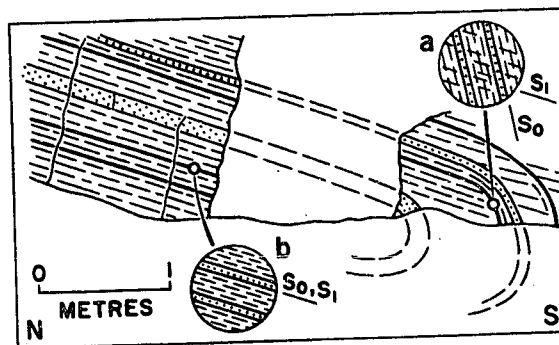


Fig. 31 Sketch of partly exposed F_1 fold in slates and siltstones. Quebec Group. Sample 25 near F_1 hinge; a) bedding S_0 , chlorite $POI\ 6.69 \pm 0.37$; axial plane foliation S_1 , $POI\ 4.16 \pm 0.04$. b) Sample 26 on limb; S_0 and S_1 parallel, $POI\ 13.40 \pm 1.20$.

non-tectonic origin, the POI of chlorite parallel to the bedding is less than 7. If the preferred orientation is the result of reorientation of the flaky minerals during burial and compaction by the overlying sediments, some such upper limit might be expected. During compaction, there is a change in volume, accompanied by escape of up to 80% of water, (Hedberg, 1936). Under these conditions the stress field may be considered axial and mineral reorientation may occur. When depth of burial increases until there is no further volume decrease, the vertical stress component rapidly approaches the horizontal stress component and hydrostatic conditions tend to prevail. The depth at which the stress field changes from axial to near hydrostatic may be about 2000 m as shales from this depth have only 5% of pore space (Athy, 1930) and little further compaction is recorded. Under such hydrostatic conditions no further mechanical mineral reorientation is to be expected.

It should be pointed out that an increase in preferred orientation of clay minerals with depth has not yet been unequivocally demonstrated, and most reorientation may take place under near-surface conditions (Meade, 1966).

In the case of a fabric produced by mineral reorientation during tectonic deformation, large variations in POI values can be expected and practically no upper limit should be found. Accordingly a large variation is illustrated by the population of POI parallel to S_0 and S_1 (Fig. 30a).

Preferred Orientation and Strain

Field observations and laboratory experiments of natural rocks and

artificial materials have demonstrated that there is a relationship between the amount of preferred orientation and strain. Williamson (1954, 1960) compared the amount of strain against the orientation of clay particles and Meade (1966) correlated compaction and preferred orientation. Clark (1970b) compared the amount of strain by compaction, flattening, and by simple shear, with an X-ray preferred orientation index and found a linear relationship between strain and preferred orientation in samples shortened from 20% to 60%. Unfortunately, the material (kaolinite-water mixture) used in these experiments had a porosity (about 50%) that is much higher than that of buried rocks undergoing deformation. Hence the results can be applied to natural rocks in only a qualitative way.

Means and Paterson (1966) compared strain with X-ray values of preferred orientation of phlogopite crystals. They used X-ray diffraction in transmission and expressed the degree of preferred orientation by the ratio of the volume of oriented material over the total volume. The material used in these experiments was a polycrystalline aggregate of phlogopite, talc, and brucite; the crystals were grown hydrothermally from their constituent oxides. Compression was 10-30% in short term experiments. When the specimens were strained at a high temperature, either during or after growth of the minerals, in all cases the process of orientation was considered to be mainly mechanical rotation after mineral growth.

The POI values of the present work are not numerically equivalent to the indices used by Means and Paterson (1966) and Clark (1970b). However, approximate conversion indicates that even at strains as high or higher than that due to natural maximum compaction, the preferred

orientation obtained experimentally is considerably less than in many samples of the chlorite along the bedding S_0 .

Two factors may explain this discrepancy:

- a) In the natural environment sedimentary and tectonic foliations form not only by mechanical reorientation of the minerals but also by recrystallization.
- b) In natural rocks preferred orientation develops by mechanical rotation of the crystals, but this process is time-dependent, therefore, in the short term experiments of Means and Paterson (1966) and Clark (1970b), the orientation of the minerals could not be fully accomplished.

Summary

The results of this investigation can be summarized as follows:

- 1) During sedimentary compaction a fabric forms which is partly dependent on lithology and possibly on depth of burial.
- 2) Approximate comparison between the X-ray values of this work and those of Means and Paterson (1966) and Clark (1970b), may indicate that in addition to mechanical reorientation during compaction, recrystallization was also operative in forming the sedimentary fabric.
- 3) The fabric parallel to the tectonic cleavages varies considerably in degree of preferred orientation. It can be lower in preferred orientation than a sedimentary fabric if the cleavage is an S_2 fracture cleavage or slaty cleavage sensu lato across the bedding; it is more highly oriented than a sedimentary fabric where the cleavage is an S_1 slaty cleavage sensu stricto parallel to the bedding. Thus the X-ray

diffraction method of determining the degree of orientation of chlorite in fine-grained rocks has quantitatively distinguished between an initial bedding-plane foliation and a tectonic foliation imposed parallel to it.

VIII SYNTHESIS

Conclusions on Structures and Fabrics

The conclusions of the investigation on folds, on cleavage types in sedimentary and low-grade metamorphic rocks, and on measurements of chlorite fabric by X-ray diffraction, have been already given in the respective chapters. In summary the main features are:

a) Fold Shape In all parts of the study area the most common types of F_1 and F_2 are class IC in the competent and 3 in the incompetent layers.

The IC folds were formed by buckling accompanied by a component of homogeneous strain (flattening), which is variable with most values no greater than 20%.

As in buckle folds measured elsewhere (Currie et al. 1962) the wavelength is almost directly a function of the thickness of the dominant member.

Contrary to expectations, there is no linear correlation between the amount of flattening and interlimb angle.

The amounts of flattening and interlimb angles for F_1 and F_2 folds are not sufficiently distinctive to use as criteria to separate fold generations. Additionally, semi-quantitative estimates throughout the study area demonstrate that the local variations in flattening and interlimb angles for F_2 folds and variations due to lithological control, are greater than the regional variation.

b) Cleavages Some cleavages observed do not fit easily into the definitions of slaty, fracture, and crenulation cleavage, and there are types

which have intermediate and other characteristics. Slaty cleavage has been subdivided into two types: sensu stricto where almost completely penetrative, and sensu lato where in distinct zones.

Lithology, amount of strain, and presence or absence of an early fabric appear to be the main factors that control cleavage style. Segregation of quartz and calcite along slaty cleavage is a common feature in almost unmetamorphosed rocks, and does not appear to be related to zones of differential shear.

F_1 and F_2 folds are characterized in rocks of similar lithology by different groups of cleavages, in particular, slaty cleavage sensu stricto appears confined to F_1 folds. However, slaty cleavage sensu lato occurs in both generations and is thus not distinctive.

c) X-ray Diffraction This technique is a useful method for measuring the preferred orientation of fine-grained phyllosilicates, which otherwise can only be estimated qualitatively. In spite of some overlapping, the X-ray values (POI) for a tectonic fabric parallel to the bedding, are significantly greater than those for a parallel fabric produced by sedimentary compaction. Thus tectonic and sedimentary foliations, though parallel, can be separated, and POI measurements can be used to detect the presence of a deformational phase in much the same manner as other structures such as fold closures.

Tectonic History

The following sequence of events is suggested for the study area:

- 1) During the Cambrian and Lower to Middle Ordovician the area was the

site of accumulation of a thick sequence of basaltic flows (Shickshock Group) interbedded with and followed by clastic sedimentary rocks (Quebec Group). The nature and thickness of the deposits (flysch and volcanics), the associated sedimentary structures (slumps etc.) all indicate a mobile environment typical of a eugeosynclinal area.

2) The first phase of regional deformation D_1 internally affected only the lower levels of the volcanic and sedimentary pile, dying upward and outward, moving from the "inner" to the "outer" part of the geosyncline (from south to north). The folds were isoclinal, probably recumbent, with schistosity and slaty cleavages virtually parallel to the bedding and metamorphic layering.

The nature and trend of F_1 folds is not well-established. However, mineral lineations and bedding-cleavage intersections (S_0/S_1) suggest that trends were approximatively east-west.

The pervasive cleavage and schistosity that are parallel to the axial planes of the folds, and the isoclinal nature of the structures indicate that a large amount of strain was involved in the deformation, which possibly was related to a zone of maximum shear at a lower level of the sedimentary pile where high temperature and pore pressure enhanced deformation (cf. Holland and Lambert, 1969, and Fyson, 1971).

Synkinematic regional metamorphism of the Shickshock and adjacent rocks of the Quebec Group accompanied this early deformation. The metamorphic grade increased southward (that is, inward in terms of geosynclinal setting) from unmetamorphosed rocks near the St. Lawrence to upper greenschist facies near the Shickshock fault.

Almost contemporaneous with the folding and regional metamorphism, a

large ultramafic body, the Mt. Albert pluton was emplaced. An increase in metamorphic grade around the pluton appears to be due to the overlapping of regional and thermal metamorphisms. Garnet porphyroblasts restricted to the surroundings of the intrusion display a synkinematic texture (curved and oblique internal fabrics), which suggests that the crystallization of garnet occurred essentially during D_1 . Field observations (S_1 foliation detected in probable early Middle Ordovician rocks), the textural evidence, and radiometric age determinations from metamorphic minerals of the Shickshock Group indicate an early Middle Ordovician age for folding, metamorphism, and emplacement of the ultramafic pluton.

3) The sedimentary and volcanic rocks were then affected, perhaps immediately following D_1 , by a second regional deformation phase D_2 . Initially there was very discontinuous folding (F_{2a}) without formation of axial-plane cleavage. Following this, F_{2b} folds formed in all the Cambro-Ordovician rocks of the study area. The folds are open to tight, upright or steeply inclined to the northwest or south, and doubly plunging.

Various axial-plane cleavages (crenulation, fracture and slaty cleavage sensu lato) were developed. Very little metamorphism accompanied the F_2 folding. Only a few muscovite and chlorite crystals grew along S_2 cleavage, mainly in the Shickshock metavolcanics and interbedded meta-sedimentary rocks, and calcite or quartz segregation occurred parallel to the cleavage.

4) A third deformational phase generated northwest trending, discontinuous upright open folds and flexures, which appear mainly in the coastal exposures of the Quebec Group. Northwest trending kink bands and chevron folds are included in this phase, also with steep axial surfaces, formed

in the foliated Shickshock and adjacent phyllitic slates of the Quebec Group.

The coastal folds and inland kink bands could have developed simultaneously, or at different times as suggested for similar structures to the southwest (Sikander, 1967). Because they were not observed in post-Ordovician rocks, they are ascribed to the last folding stage of the Taconic orogeny.

5) Following uplift, erosion, and subsidence, the sedimentation of well-sorted sandstone and limestone during the Siluro-Devonian was typical of a platform environment, but the thickness of the succession (more than 6 km, is much greater than usual for platform deposits).

6) The Silurian and Devonian rocks were gently folded during the mid-Devonian Acadian orogeny. Also formed at this time were large-scale, east-northeast trending folds in Cambro-Ordovician rocks, notably the Tourelle syncline and the Shickshock anticline. Granitic plutons (Table Top granite) were emplaced at the end of the Acadian orogeny.

Structural History of Lower Paleozoic rocks, Gaspé Peninsula

An attempt was made to correlate structures observed in several areas, mainly in North Gaspé, to see if the deformational phases of the study area represent local tectonic events or if they are present throughout the region.

In Table 4 the successions of structures in Cambro-Ordovician rocks are summarized for three areas, Témiscouata in the southwest, Matane and Mt. Albert in the northeast (Fig. 32). The similarity of F_1 structures

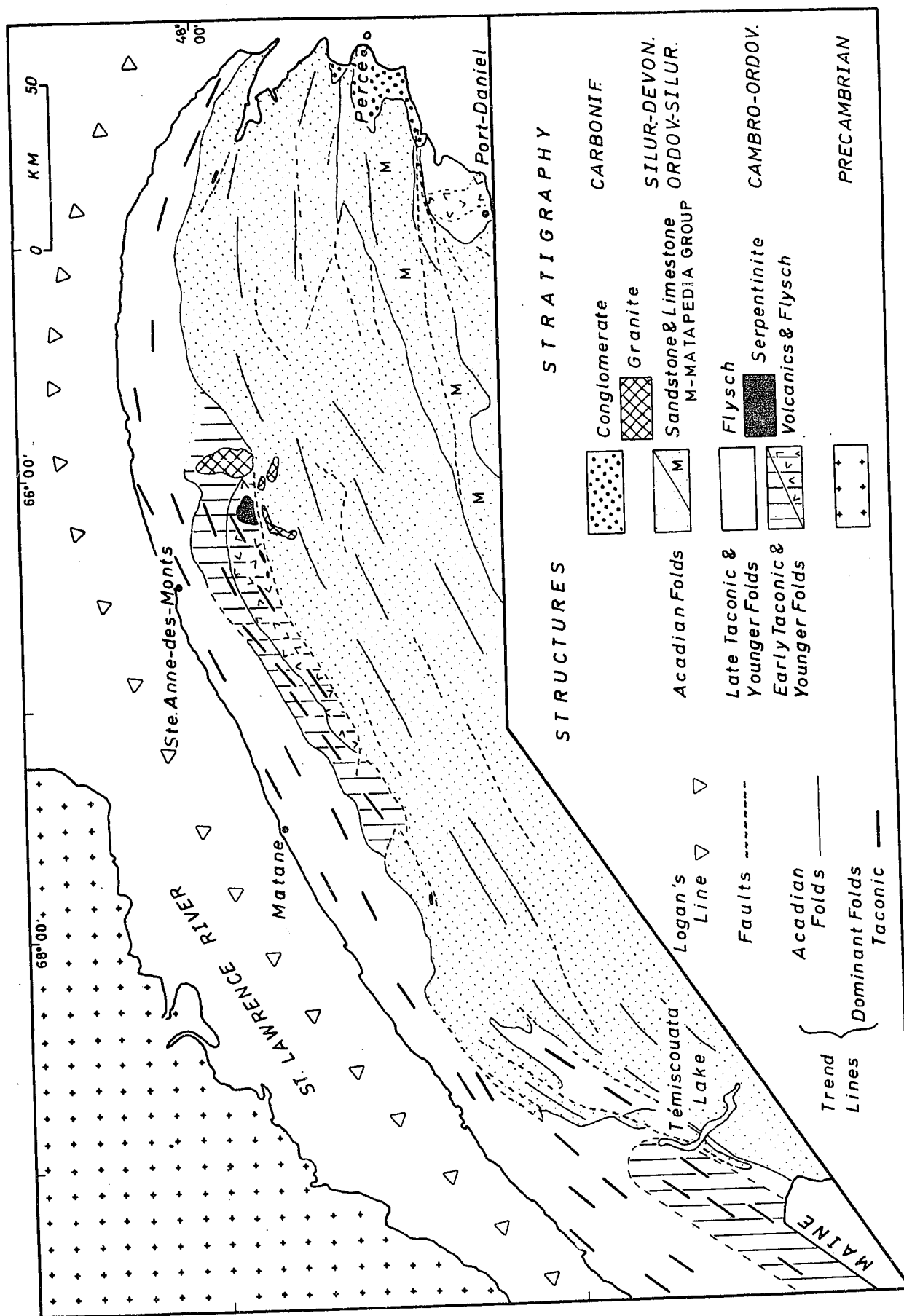


Fig. 32 Areal distribution of the Taconic and Acadian major structures in Gaspé Peninsula. Geology modified after geological map of Quebec Department of Mines, 1969, (Laurin).

in each area is apparent, and thus these are assigned to a single regional event of deformation, Dr_1 . This does not imply that the Dr_1 deformation was synchronous throughout the region, but merely that within a given time interval (somewhere within the Middle or Upper Ordovician) this type of deformation was taking place. The F_1 isoclinal folds are demonstrated to occur in a narrow discontinuous belt along the peninsula (Fig. 32), mainly, as in the study area, by identifying the presence of a tectonic foliation parallel to bedding.

F_2 folds affect all the Cambro-Ordovician rocks in the region, and as they are areally continuous, they are readily assigned to a Dr_2 deformation. The upper limit to the age of Dr_2 deformation is defined in the Témiscouata area by the presence of Lower Silurian rocks lying unconformably on F_2 folds. Silurian rocks also unconformably overlie these folds at Cap-des-Rosières at the eastern tip of the peninsula. Sikander and Fyson (1969), however, referred the F_2 folds in the Matane area to the Acadian orogeny as here they are parallel in trend to folds in Silurian and Devonian rocks that are in fault contact. But the stratigraphic evidence outside the Matane area appears to be more reliable in giving a general age than the structural similarities (see also Appendix IV).

The correlation into regional phases of deformation of the areally restricted folds (F_3 in Mt. Albert area, F_3 , F_4 in the Matane area, and F_3 , F_4 in Témiscouata area) is difficult. However, in all these areas there are northwesterly trending open folds or kink bands, which appear to be among the last small-scale folds formed during the Taconic orogeny; here they are assigned to a Dr_4 phase of regional deformation.

A shallow-dipping crenulation cleavage, axial plane to small open

Table 4 Comparison of Structural Successions in Three Areas of Cambro-Ordovician Rocks, North Gaspé

Regional Succession	Fold Character	Témiscouata Area *	Matane Area **	Mt. Albert Area *
		open-gentle (F ₅)		open-gentle (F ₄)
Dr ₅	Orientation	NNE upright		ENE upright
	Cleavage	-		-
	Metamorphism	-		contact
	Magmatism	-		acid intrusions
	Age	mid-Devonian		mid-Devonian
Dr ₄ ***	Fold Character	? (F ₄)	open (F ₃) & kinks (F ₄)	open & kinks (F ₃)
	Orientation	NW upright	E-W upright (F ₃) NW upright (F ₄)	NW upright
	Cleavage	fracture & crenulation	fracture	fracture
	Metamorphism	-	-	-
	Magmatism	-	-	-
	Age	Upper (?) Ordov.	Upper (?) Ordov.	Upper (?) Ordov.
Dr ₃ ***	Fold Character	open (F ₃)	micro-kinks	
	Orientation	recumbent	recumbent	
	Cleavage	crenulation	-	
	Metamorphism	-	-	
	Magmatism	-	-	
	Age	Ordovician (?)	Ordovician or Devonian	
Dr ₂	Fold Character	tight-open (F ₂)	tight-open (F ₂)	tight-open (F _{2a} , F _{2b})
	Orientation	NNE upright	NE upright	NE upright
	Cleavage	crenulation & fracture	crenulation & fracture	crenulation & fracture
	Metamorphism	-	(?)	little muscovite-chlorite growth
	Magmatism	-	-	-
	Age	Middle or Upper Ordovician		Middle or Upper Ordovician
Dr ₁	Fold Character	isoclinal (F ₁)	isoclinal (F ₁)	isoclinal (F ₁)
	Orientation	NE (?) recumbent	NE (?) recumbent	E-W (?) recumbent
	Cleavage	slaty & schistosity	schistosity & slaty	schistosity & slaty
	Metamorphism	to lower green-schist facies	to upper green-schist facies	to upper greenschist facies
	Magmatism	-	-	ultramafic intrusions
	Age	Middle or Upper Ordovician	Middle or Upper Ordovician	early Middle Ordovician

* Areas studied by the author

** Area studied by Sikander (1967), and Sikander and Fyson (1969); age of structural events modified by the author.

*** Structures produced are sporadically distributed.

Dr - Regional succession of deformational phases for Gaspé Peninsula.

F₁, F₂, etc. - Local succession of folding for each area studied.

Age - Designates age determined from evidence within area.

recumbent folds, occurs locally in phyllitic rocks of the Quebec Group in the Témiscouata area; here these structures are assigned to an F_3 phase of deformation because the horizontal cleavage is cut by the northwesterly trending vertical cleavage (F_4). However, Sikander and Fyson (1969) observed low-dipping kink bands in both Cambro-Ordovician and Siluro-Devonian rocks of the Matane area, and they ascribed all these structures to a late-Acadian areally discontinuous phase. Because in the Témiscouata area the F_3 structures of similar orientation appear to be Taconic features, in the Matane area the low-dipping kink bands in the Cambro-Ordovician rocks may be later than those in the Témiscouata area.

Large-scale gentle folds (F_5), defined solely by the outcrop pattern with no small-scale equivalents, are apparent in the Ordovician rocks of the Mt. Albert and Témiscouata areas. In both areas these folds are considered as Acadian structures due to the conformity with similar folds in nearby Devonian rocks, including (at Témiscouata) folds along the regional trend (see also Appendix IV). Thus the Acadian folding appears to have affected the Cambro-Ordovician "basement". The lack of stratigraphic control within the Cambro-Ordovician rocks of the Matane area, hinders the identification of similar features.

More tentative is the correlation of the structural succession of North Gaspé with other parts of the peninsula. However, in the Port-Daniel area to the south of the Gaspé synclinorium (Fig. 32), Ayrton (1964, 1967) recognized three main tectonic phases on the basis of fold generations and unconformities. The first folding, affecting the clastic sediments and mildly metamorphosed volcanics of the Maquereau Group, was referred to a pre-Middle Ordovician "Gaspesian orogeny". However, the

folds could be an expression of D_1 of North Gaspé. Unlike in North Gaspé, the stratigraphic upper boundary to the F_1 folding appears to be marked by an unconformity, and Ayrton (1967) noted fragments of Maquereau rocks with internal cleavages in the basal (?) conglomerate of the overlying Middle Ordovician Mictaw Group. The shales and greywackes of the Mictaw Group are affected by folds of similar form to F_2 folds in north Gaspé, and they underlie Silurian rocks unconformably. Although northern Gaspé and Port-Daniel area appear to have experienced a similar Taconic structural history, in the Matapedia Group of south central Gaspé (Fig. 32), Ordovician and Silurian sediments comprise a conformable sequence which was deformed extensively only during the Acadian orogeny (Béland, 1969). The lack of the Taconic unconformity in the Matapedia Group suggests that the Taconic orogeny was essentially an areally discontinuous event (Pavlidies et al. 1968).

In conclusion, the structural history of Gaspé Peninsula insofar as it affected Cambro-Ordovician rocks can be summarized as follows:

- 1) Middle Ordovician severe folding, mild metamorphism, and emplacement of ultrabasic bodies occurred along a narrow discontinuous (?) northern belt, and similar deformation and metamorphism affected some localities in southern Gaspé. Deformation affected only the lower levels of the sedimentary sequence, and its upper limit was not marked in northern Gaspé by an angular unconformity.
- 2) Further deformation was expressed as widespread then discontinuous folding. This was followed by uplift and erosion in North Gaspé and the Port-Daniel area, but not in the Matapedia belt (Fig. 32), where the rocks were not deformed during the Taconic orogeny.

- 3) Acadian folding, pronounced in the Matapedia belt in the south, faded out to the north (Béland, 1969), where only gentle warping affected the previously deformed Cambro-Ordovician rocks on the northern flank of the Gaspé synclinorium. Nevertheless the general outcrop pattern in the Cambro-Ordovician rocks is determined by these structures.
- 4) The emplacement of acid plutons in the "foreland" of the Acadian geosyncline through Cambro-Ordovician and younger rocks marked the end of Devonian tectonism.
- 5) Faulting occurred in part during the Carboniferous (Rodgers, 1971).

Tectonic Speculations

Near Percé, Middle Ordovician muddy limestone rests unconformably on shelf-type carbonate rocks of Cambrian age (Fritz et al. 1970). The basal rocks are of a totally different facies from the Cambro-Ordovician Quebec Group and the Maquereau Group of Port-Daniel. The facies relations might support the hypothesis that all the northern belt of Gaspé is a huge allochthon, which slid northward onto the carbonate platform of the foreland of the Taconic geosyncline, as exposed on Anticosti Island (Poole, 1967, Rodgers, 1971). The Cambrian carbonates of Percé thus may be a fragment of the platform largely buried by allochthonous geosynclinal deposits.

The above hypothesis is not substantiated at present by structural data and more work is needed before it is accepted or modified.

Along similar lines, Bird and Dewey (1970) suggested that the Cloridorme Formation represents a tectonic slide coming from the south.

However, the contact between the Cloridorme rocks and the underlying Quebec Group is gradational and in some localities stratigraphically interfingering.

Apart from the problem of the autochthonous or allochthonous nature of the Cambro-Ordovician rocks of North Gaspé, it is of interest to consider the relationships between the axes of the Taconic and later Acadian folds (Fig. 32). The Acadian (Dr_5) folds are generally parallel with the dominant (Dr_2) Taconic folds. However, locally as at Mt. Albert and in the northern part of the Témiscouata area, the axial traces of the two sets are at angles varying from a few degrees up to almost 40° .

The general parallelism in trend of folds of different ages can be ascribed to an "edge effect", that is the Acadian trends were controlled by the orientation of the Taconic folds in the bordering and, at least in part, underlying basement. In a similar fashion the arcuate shape of Gaspé Peninsula is probably due to the original shape of the geosynclinal trough which was controlled by the edge of the Precambrian craton.

Regarding the relationships between the tectonic history of Gaspé Peninsula and plate-tectonic theory, evidence for or against the hypothesis that plate processes were operative in the structural development of Gaspé, is scanty. The linear belt of volcanic rocks (Shickshock Group) and ultramafic bodies (Fig. 32) along the southern border of the Cambro-Ordovician sedimentary rocks of the Quebec Group may indicate the presence of oceanic crust and of a subduction zone (Bird and Dewey, 1970). However, buoyant bodies rising from the mantle (Ramberg, 1971) can also account for such features.

IX SUGGESTIONS FOR FUTURE WORK

Future structural work in the Gaspé area could usefully proceed along the following lines:

- a) Collection of a large number of folded layers and study of as many morphological features as possible by using an Istronic Graticon Digitizer (now available in many research institutes). In this way, curvature, layer thickness, interlimb angle, wavelength, etc. can be recorded on computer cards. This enables a complete and fast treatment of a large, statistically significant number of data without tedious and time-consuming manual work.
- b) By further X-ray work, possibly by means of a pole-goniometer, analyze the orientation of phyllosilicates on cleavage planes. The quantitative measurements of fabric so obtained can be used for various purposes such as differentiation between sedimentary and tectonic foliations, study of the cleavage in folds of different shapes (open, close, or with rounded or sharp hinges, etc.) and in different parts of the same fold, and possibly establishing cleavage types that are distinctive for various fold generations. In particular numeric values of cleavage might be plotted against the approximate total amount of strain as indicated by the fold shape.
- c) Use the electron microscope to study textures in very fine grained sedimentary rocks. From the X-ray work integrated with the electron microscope and standard optical investigation, it may be possible to improve the understanding of the nature of cleavage.
- d) Further radiogenic age determinations on the metamorphic minerals of

the Shickshock and possibly of the Quebec Group, associated with more detailed stratigraphic work; this could be useful in determining more precisely the age of F_1 and F_2 folding.

e) Further structural work in the Témiscouata area and to the southwest. In this way it might be possible to correlate the structural history of North Gaspé with the complex tectonic events that occurred in the Eastern Townships. For this purpose, F_2 folds, which are widespread features can be traced along the St. Lawrence shore southwestward and used as time markers to which other deformations can be referred.

REFERENCES

- Alcock, F.J., 1924, Shickshock Mountains, Central Gaspé, Quebec: Geol. Surv. Canada, Summary Rept., Pt. C, p. 127-133.
- _____ 1926, Mount Albert map area, Quebec: Geol. Surv. Canada Mem. 144, 75 p.
- Athy, L.F., 1930, Density, porosity, and compaction of sedimentary rocks: Amer. Assoc. Petroleum Geologists Bull., v. 14, p. 1-24.
- Ayrton, W.G., 1964, A structural study of the Chandler-Port Daniel area, Gaspé peninsula, Quebec: Ph. D. thesis, Northwestern University, Illinois.
- _____ 1967, Chandler-Port Daniel area: Quebec Dept. Nat. Res. Geol. Rept. 120, 91 p.
- Baer, A., 1956, La schistosité et sa répartition, résumé des recherches des M.P. Fourmarier: Geol. Rundschau, v. 45, p. 234-247.
- Baker, D.W., Wenk, H.R., and Christie, J.M., 1969, X-Ray analysis of preferred orientation in fine-grained quartz aggregates: Jour. Geology, v. 77, p. 144-172.
- Beavis, F.C., 1970, Cleavage and linear structures in Ordovician sandstones, Victoria: Geol. Mag. v. 107, p. 429-438.
- Béland, J., 1957, Ste. Felicite-Grosses Roches area, Gaspé, Quebec: Quebec Dept. Mines, Prelim. Rept. 339, 8 p.
- _____ 1960, Rimouski-Matapedia area: Quebec Dept. Mines, Prelim. Rept., 430, 18 p.
- _____ 1967, Contributions from systematic studies of minor

- structures in the southern Quebec Appalachians: Roy. Soc. Canada Spec. Pub. 10, p. 48-56.
- _____ 1969, The Geology of Gaspé: Can. Inst. Mines Trans., v. LXXII, p. 213-220.
- Billings, M.P., 1954, Structural Geology: Prentice-Hall Inc., Englewood Cliffs, U.S.A., 514 p.
- Biot, M.A., 1957, Folding instability of a layered viscoelastic medium under compression: Proc. Roy. Soc. London, Ser. A, v. 242, p. 444-454.
- _____ 1961, Theory of folding of stratified viscoelastic media and its implication in Tectonics and Orogenesis: Geol. Soc. America Bull., v. 72, p. 1595-1620.
- _____ 1964, Theory of internal buckling of a confined multilayered structure: Geol. Soc. Amer. Bull., v. 75, p. 563-568.
- _____, Odé, H. and Roever, W.L., 1961, Experimental verification of the folding of stratified viscoelastic media: Geol. Soc. America Bull., v. 72, p. 1621-1630.
- Bird, J.M., and Dewey, J.F., 1970, Lithosphere plate-continental margin Tectonics and the evolution of the Appalachian orogen: Geol. Soc. America Bull., v. 81, p. 1031-1060.
- Braddock, W.A., 1970, The origin of slaty cleavage-evidence from Precambrian rocks in Colorado: Geol. Soc. America Bull., v. 81, p. 589-600.
- Carrara, A., and Fyson, W.K., 1971, Comparison of sedimentary and tectonic foliations by X-ray diffraction: Can. Jour. Earth Sci., v. 8, p. 1190-1196.

- Chapple, W.M., 1968, A mathematical theory of finite-amplitude rock folding: Geol. Soc. America Bull., v. 79, p. 47-68.
- Clark, D.R., 1970a, Mechanical formation of preferred orientation in clays: Amer. Jour. Sci., v. 269, p. 250-266.
- _____ 1970b, Origin of slaty cleavage in the Coeur d'Alene district, Idaho: Geol. Soc. America Bull., v. 81, p. 3061-3072.
- Cloos, E., 1947, Oolite deformation in the South Mountain fold, Maryland: Geol. Soc. America Bull., v. 58, p. 843-918.
- Crickmay, G.W., 1932, Evidence of Taconic Orogeny in Matapedia Valley: Amer. Jour. Sci., v. 24, p. 368-386.
- Crook, K.A.W., 1964, Cleavage in weakly deformed mudstones: Amer. Jour. Sci., v. 262, p. 523-531.
- Cumming, L.M., 1961, Silurian and Lower Devonian formations in the eastern part of the Gaspé Peninsula, Quebec: Geol. Surv. Canada Mem. 304, 45 p.
- Currie, J.B., Patnode, H.W., and Trump, R.P., 1962, Development of folds in sedimentary strata: Geol. Soc. America Bull., v. 73, p. 655-674.
- Dawson, C.G., 1927, Een eenvoudige toestand om de orientatie en de structuur van kristallen met behulp van X-strahlen te bepalen: Physica, v. 7, p. 302-314.
- de Römer, H.S., 1970, Madeleine Nord River area, Gaspé North County: Quebec Dept. Nat. Res. Prelim. Rept. 594, 19 p.
- de Sitter, L.V., 1964, Structural Geology: McGraw Hill Book Co., New York, 551 p.
- Decker, B.F., Asp, E.T., and Harker, D., 1943, Preferred orientation

- determination using a Geiger-counter X-ray diffraction goniometer: Jour. Applied Physics, v. 19, p. 388-392.
- den Tex, E., 1963, A commentary on the correlation of metamorphism and deformation in space and time: Nederlands Geol. Mijnb., v. 42, p. 170-176.
- Dieterich, J.H., 1969, Origin of cleavage in folded rocks: Amer. Jour. Sci., v. 267, p. 155-165.
- Divi, R., 1972, Structural analysis of Grenville rocks east of Bancroft, Ontario, Canada: Ph. D. thesis, University of Ottawa, Ontario.
- Donath, F.A., and Parker, R.B., 1964, Folds and folding: Geol. Soc. America Bull. v. 75, p. 45-62.
- Ells, R.W., 1885, Report on exploration and surveys in the interior of Gaspé peninsula (1883) Geol. Surv. Canada, Prog. Rept. 1882-84, p. 1-34.
- Engelhardt, W., and Gaida, G.H., 1963, Concentration changes of pore solutions during the compaction of clay sediments: Jour. Sed. Petrology, v. 33, p. 919-930.
- Engels, B., 1959, Über neue Ergebnisse kleintektonischer Untersuchungen Schiefer-gebirge: Geol. Rundschau, v. 48, p. 271-280.
- Enos, P., 1969, Cloridorme Formation, Middle Ordovician flysch, northern Gaspé Peninsula, Quebec: Geol. Soc. America Spec. Paper 117, 66 p.
- Fairbairn, H.W., 1943, X-ray petrology of some fine-grained foliated rocks: Amer. Mineralogist, v. 28, p. 246-256.
- Field, M., and Marchant, M.E., 1949, Reflection method of determining preferred orientation on the Geiger counter spectrometer: Jour.

Applied Physics, v. 20, p. 741-745.

Fritz, W.H., Kindle, C.H., and Lespérance, P.J., 1970, Trilobites and stratigraphy of the Middle Cambrian Corner-of-the Beach Formation, eastern Gaspé Peninsula, Quebec: Geol. Surv. Canada Bull. 187, p. 43-58.

Fyson, W.K., 1962, Tectonic structures in the Devonian rocks near Plymouth Devon: Geol. Mag., v. 99, p. 208-226.

_____ 1971, Fold attitudes in metamorphic rocks: Amer. Jour. Sci. v. 270, p. 373-382.

Girard, P., 1967, Mount Richardson area: Quebec Dept. Nat. Res., Prelim. Rept. 563, 11 p.

_____ 1969, Geology of Madeleine Mines Deposit: Can. Inst. Mines Trans., v. LXXII, p. 239-247.

_____ 1971, The Madeleine Copper Mine, Gaspé, Quebec: a hydrothermal deposit: Ph. D. thesis, McGill University, Quebec.

Hedberg, H.D., 1936, Gravitational compaction of clay and shales: Amer. Jour. Sci. v. 31, p. 241-297.

Higgs, D.V., Friedman, M. and Gebhardt, J.E., 1960, Petrofabric analysis by means of the X-rays diffractometer in Griggs, D.V., and Handin, J., editors Rock Deformation (a symposium): Geol. Soc. America Mem., v. 79, p. 275-292.

Ho, T.L., 1947, Petrofabric analysis by X-rays: Geol. Soc. China Bull., v. 27, p. 389-398.

Hobbs, B.E., 1971, The analysis of strain in folded layers: Tectonophysics, v. 11, p. 329-375.

Holland, J.G., and Lambert, R. St. J., 1969, Structural regimes and

- metamorphic facies: Tectonophysics, v. 7, p. 197-217.
- Jetter, L.K., and Borie, B.S., Jr., 1953, Method for the quantitative determination of preferred orientation: Jour. Applied Physics, v. 24, p. 532-535.
- Jones, I.W., 1933, Marsoui Map-area, Gaspé peninsula: Quebec Dept. Mines Ann. Rept. 1932, Pt. D, p. 3-32.
- Kaarsberg, E.A., 1959, Introductory studies of natural and artificial argillaceous aggregates by sound-propagation and X-ray diffraction methods: Jour. Geology, v. 67, p. 447-472.
- Kamb, W.B., 1959, Ice petrofabric observations from Blue Glacier, Washington, in relation to theory and experiment: Jour. Geophys. Res., v. 64, p. 1891-1909.
- Knill, J.L., 1960, A classification of cleavages, with special references to the Craginsh district of the Scottish Highlands: Inter. Geol. Cong. XXI Copenhagen, Pt. 18, p. 317-325.
- _____ 1961, Joint-drags in Mid-Argyllshire: Proc. Geol. Assoc., v. 72, p. 13-19.
- Kratky, K.O., 1930, Ein Roentgengoniometer für die Polykristalluntersuchung: Zeitsschr. Kristallographie, v. 72, p. 529.
- Lajoie, J., 1961, Lac Prime West, and part of Wild Goose area: Quebec Dept. Nat. Res. Prelim. Rept. 448, 9 p.
- _____ 1962, Chénier-Bédard (Lac-des-Baies) area: Quebec Dept. Nat. Res. Prelim. Rept. 493, 8 p.
- _____, Lespérance, P.L. and Béland, J., 1968, Silurian stratigraphy and paleogeography of Matapedia-Témiscouata region, Quebec: Amer. Assoc. Petroleum Geologists Bull., v. 52, p. 615-640.

- Leech, G.B., Lowdon, J.A., Stockwell, C.H., and Wanless, R.K., 1963, Age determination and geological studies: Geol. Surv. Canada Paper 63-17, 140 p.
- Leith, C.K., 1923, Structural Geology (2nd edition): Henry Holt and Company, New York, 390 p.
- Lespérance, P.J., and Greiner, 1969, Squateck-Cabano area: Quebec Dept. Nat. Res. Geol. Rept. 128, 111 p.
- Logan, W.E., 1846, Report of progress for the year 1844: Geol. Surv. Canada, Rept. Prog. 1844, 110 p.
- _____ 1863, Geology of Canada: Geol. Surv. Canada Rept. Prog. to 1863, 983 p.
- Loudon, T.V., 1964, Computer analysis of orientation data in structural geology: Office Naval Res., Geograph. Branch Tech. Rept. 13, 130 p.
- Low, A.P., 1885, Report on exploration and surveys of the interior of the Gaspé peninsula: Geol. Surv. Canada Rept. Prog. 1882-84, 21 p.
- Lowdon, J.A., Stockwell, C.H., Tipper, W.H., and Wanless, R.K., 1963, Age determinations and geological studies: Geol. Surv. Canada Paper 62-17, 140 p.
- Martin, R.T., 1966, Quantitative fabric of wet kaolinite: Mass. Inst. Technology, Soil. Engr. Div. Pub. 127, 28 p.
- Matte, P., 1969, Le problème du passage de la schistosité horizontale à la schistosité verticale dans le dôme de la Garonne (Paléozoïque des Pyrénées Centrales): Acad. sci. Paris Comptes rendus, Sér. D, v. 268, p. 1841-1844.
- Mattinson, C.R., 1958, The geology of the Mount Logan area, Gaspé, Quebec: Ph. D. thesis, McGill University, Quebec.

- _____ 1964, Mount Logan area, Gaspé peninsula: Quebec Dept. Mines, Geol. Rept. 118, 97 p.
- MacGregor, I.D., 1962, Geology, petrology and geochemistry of the Mount Albert and associated ultramafic bodies of Central Gaspé, Quebec: M. Sc. thesis, Queen's University, Ontario.
- Maxwell, J.C., 1962, Origin of slaty and fracture cleavage in the Delaware Water Gap area, New Jersey and Pennsylvania, in Engel, A.E.J., James, H.L., and Leonard, B.F., editors, Petrologic Studies: A volume in honor of A.E. Buddington: Geol. Soc. America, p. 281-311.
- McBirney, A.R., and Best, M.G., 1961, Experimental deformation of viscous layers in oblique stress fields: Geol. Soc. America Bull., v. 72, p. 495-498.
- McGerrigle, H.W., 1950, The geology of eastern Gaspé: Quebec Dept. Mines, Geol. Rept. 35, 168 p.
- _____ 1953, Geological map of the Gaspé Peninsula: Quebec Dept. Mines, map 1000.
- _____ 1954, The Tourelle and Courcellette areas: Quebec Dept. Mines, Geol. Rept. 62, 63 p.
- _____ and Skidmore, W.B., 1967, Geological map of Gaspé Peninsula: Quebec Dept. Nat. Res., map 1642.
- Meade, R.H., 1961, X-ray diffractometer method for measuring preferred orientation in clays: U.S. Geol. Surv. Paper 424-B, p. B273-B276.
- _____ 1966, Factors influencing the early stages of the compaction of clays and sands-review: Jour. Sed. Petrology, v. 36, p. 1085-1101.

- Means, W.D., 1963, Mesoscopic structures and multiple deformation in the Otago schists: New Zealand Jour. Geol. Geophy., v. 6, p. 801-816.
- _____ 1966, A macroscopic recumbent fold in schists near Alexandra, Central Otago: New Zealand Jour. Geol. Geophy. v. 9, p. 173-194.
- _____ and Paterson, M.S., 1966, Experiments on preferred orientation of platy minerals: Contr. Mineral. Petrol. v. 13, p. 108-133.
- Moench, R.H., 1966, Relation of S_2 schistosity to metamorphosed clastic dikes, Rangeley-Phillips area, Maine: Geol. Soc. America Bull., v. 77, p. 1449-1462.
- Murray, A., 1847, Report on the geology of a portion of Gaspé, including the Matane, Ste-Anne, and St. John Rivers: Geol. Surv. Canada Rept. Prog. 1845-46, p. 99-118.
- Oertel, G., 1970, Deformation of a slaty, lapillar tuff in the Lake district, England: Geol. Soc. America Bull., v. 81, p. 1173-1188.
- Ollerenshaw, N.C., 1961, Couq-Langis area: Quebec Dept. Mines, Prelim. Rept. 465, 14 p.
- _____ 1963, Stratigraphic problem of the western Shickshock Mountain in Gaspé peninsula: Ph. D. thesis, University of Toronto, Ontario.
- _____ 1967, Couq-Langis area: Dept. Mines, Geol. Rept. 121, 192 p.
- Osberg, P.H., 1965, Structural geology of the Knowlton-Richmond area,

- Quebec: Geol. Soc. America Bull., v. 76, p. 223-250.
- Park, R.G., 1969, Structural correlation in metamorphic belts: Tectonophysics, v. 7, p. 323-338.
- Pavlides, L., Boucot, A.J., and Skidmore, W.D., 1968, Stratigraphic evidence for the Taconic orogeny in the northern Appalachians: in Zen, E-an, White, W.S., Hardley, J.B., and Thompson Jr., J.B., editors, Studies of Appalachian geology, northern and maritime: p. 61-82.
- Poole, W.H., 1967, Tectonic evolution of Appalachian region of Canada: Geol. Assoc. Canada Spec. Paper 4, p. 9-51.
- Powell, C. McA., 1969, Intrusive sandstone dykes in the Siamo slate near Negaunee, Michigan: Geol. Soc. America Bull., v. 80, p. 2585-2594.
- Quigley, R.M., and Thompson, C.D., 1966, The fabric of anisotropically consolidated sensitive marine clay: Can. Geotech. Jour., v. 3, p. 61-73.
- Ramberg, H., 1952, Origin of Metamorphic and Metasomatic Rocks: Univ. Chicago Press, Chicago, 317 p.
- _____ 1959, Evolution of ptygmatic folding: Norsk Geol. Tidsskr., v. 39, p. 99-151.
- _____ 1960, Relationships between length of arc and thickness of ptygmatically folded veins: Amer. Jour. Sci., v. 258, p. 36-46.
- _____ 1961, Contact strain and folding instability of a multilayered body under compression: Geol. Rundschau, v. 51, p. 405-439.
- _____ 1963, Strain distribution and geometry of folds: Bull. Geol. Inst. Univ. Upsala, v. 42, p. 1-20.

- _____ 1964, Selective buckling of composite layers with contrasted rheological properties; a theory for simultaneous formation of several orders of folds: *Tectonophysics*, v. 1, p. 307-341.
- _____ 1971, Dynamic models simulating rift valleys and continental drift: *Lithos*, v. 4, p. 259-276.
- _____ and Ghosh, S.K., 1968, Deformation structures in the Hovin Group schists in the Hommelvic-Hell region (Norway): *Tectonophysics*, v. 6, p. 311-330.
- _____ and Strömberg, K.E., 1971, Experimental tests of modern buckling theory applied on multilayered media: *Tectonophysics*, v. 11, p. 461-472.
- Ramsay, J.G., 1962, The geometry and mechanics of formation of "similar" type folds: *Jour. Geology*, v. 70, p. 309-327.
- _____ 1967, *Folding and fracturing of rocks*: McGraw Hill Book Company, New York, 568 p.
- _____ and Graham, R.H., 1970, Strain variation in shear belts: *Can. Jour. Earth Sci.* v. 7, p. 786-813.
- Reesor, J.E., 1970, Some aspects of structural evolution and regional setting in part of the Shuswap Metamorphic Complex, in Wheeler, J.O., editor, *Structures of the Southern Canadian Cordillera*: *Geol. Assoc. Canada Spec. Paper* 6, p. 73-86.
- Richardson, J., 1859, Report of Mr. James Richardson, explorer (on portion of Gaspé peninsula): *Geol. Surv. Canada Rept. Prog.* p. 105-169.
- Rickard, M.J., 1961, A note on cleavages in crenulated rocks: v. 98, p. 324-332.

- Riva, J., 1968, Graptolite faunas from the Middle Ordovician of the Gaspé north shore: *Natur. Can.*, v. 95, p. 1379-1400.
- _____, 1969, Middle and Upper Ordovician graptolite faunas of St. Lawrence Lowlands of Quebec, and of Anticosti Island: *Amer. Assoc. Petroleum Geologists Mem.* 12, p. 513-556.
- Robert, J.L., 1966, Mount Hog's Back area: Quebec Dept. Nat. Res. Prelim. Rept. 540, 25 p.
- Rodgers, J., 1971, The Taconic orogeny: *Geol. Soc. America Bull.*, v. 82, p. 1141-1178.
- _____, and Neale, E.R.W., 1963, Possible Taconic klippen in western Newfoundland: *Am. Jour. Sci.* v. 261, p. 713-730.
- Rust, B.R., 1965, The stratigraphy and structures of the Whithorn area of Wigtownshire, Scotland: *Scottish Jour. Geol.*, v. 1, p. 101-133.
- Sander, B., and Sachs, G., 1930, Zur Röntgenoptischen Gefügeanalyse von Gesteinen: *Zeitschr. Kristallographie*, v. 75, p. 550-571.
- Schmidt, W., 1932, Tektonik und Verformungslehre: Gebr. Borntraeger, Berlin, 208 p.
- Schultz, L.G., 1949, A direct method of determining preferred orientation of a flat sample using a Geiger counter X-ray spectrometer: *Jour. Applied Physics*, v. 20, p. 1030-1033.
- Schwerdtner, W.M., Sheehan, P.M. and Rucklidge, J.C., 1971, Variation in degree of hornblende grain alignment within two boudinage structures: *Can. Jour. Earth Sci.*, v. 8, p. 144-149.
- Sherwin, J.A., and Chapple, W.M., 1968, Wavelengths of single layer folds - a comparison between theory and observation: *Amer. Jour. Sci.*, v. 266, p. 167-179.

- Sikander, A.H., 1967, Structural analysis of the Lower Paleozoic rocks of western Gaspé, Quebec: Ph. D. thesis, Ottawa University, Ontario.
- _____ and Fyson, W.K., 1969, Structural development of the Paleozoic rocks of western Gaspé, Quebec: Can. Jour. Earth Sci., v. 6, p. 1113-1127.
- Silverman, E.N., and Bates, T.F., 1960, X-ray diffraction study of orientation in the Chattanooga shale: Amer. Miner., v. 45, p. 60-68.
- Spry, A., 1969, Metamorphic Textures: Pergamon Press, Oxford, 350 p.
- St. Julien, P. 1967, Tectonics of part of Appalachian region of southeastern Quebec (southwest of the Chaudière River): Roy. Soc. Canada Spec. Pub. 10, p. 41-47.
- Stabler, C.L., 1968, Simplified Fourier analysis of fold shapes: Tectonophysics, v. 6, p. 343-350.
- Talbot, J.L., and Hobbs, B.E., 1968, The relationship of metamorphic differentiation to other structural features at three localities: Jour. Geol., v. 76, p. 581-587.
- Tanner, J.G. and Uffen, R.J., 1960, Gravity anomalies in the Gaspé Peninsula, Quebec: Publication of Canada Dept. Mines and Tech. Surveys Dom. Obs., v. 21, no. 5, p. 221-260.
- Turner, F.J., 1941, The development of pseudo-stratification by metamorphic differentiation in the schists of Otago, New Zealand: Amer. Jour. Sci., v. 239, p. 1-16.
- _____ and Weiss, L.E., 1963, Structural Analysis of Metamorphic Tectonites: McGraw-Hill Book Co., New York, 545 p.

Waver, F., 1924, Über die Walzstruktur kubisch kristallisierender Metalle:

Physik, v. 28, 69 p.

Williams, P.F., 1970, A criticism of the use of style in the study of deformed rocks: Geol. Soc. America Bull., v. 81, p. 3283-3296.

_____ 1972, Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia: Amer. Jour. Sci., v. 272, p. 1-47.

Williamson, W.O., 1954, The effects of rotational rolling on the fabric and dry shrinkage of clay: Amer. Jour. Sci. v. 252, p. 129-143.

_____ 1960, Some effects of deformation on the structure and properties of clay, in Mineral Industries: Penn. State Univ., v. 29, p. 1-5.

APPENDIX I

K-Ar Ages of the Samples from Mount Albert Area

Sample	Mineral	Percent K ₂ O	Percent Radiogenic Ar-40	Ar ⁴⁰ /K ⁴⁰	Age in m.y.
III-343	Biotite	6.23	97	0.0312	469±26 ^o
" "	" "	---	91	0.0303	460±25
III-344	Biotite	5.22	oo	0.0290	440±24
III-469	Hornblende	0.43	63	0.0315	474±45

Potassium and argon were made on aliquots of 60-35 mesh samples. Potassium was determined by flame photometry. Argon was determined on an A.E.I. MS-10 mass spectrometer operated statically using argon - 38 as a spike. The following constants were used in the calculations:

$$\begin{aligned}\lambda_e &= 0.0585 \times 10^{-10} \text{ yr}^{-1} \\ \lambda &= 5.30 \times 10^{-10} \text{ yr}^{-1} \\ K^{40}/K &= 0.0119 \text{ atom percent}\end{aligned}$$

The above analyses were carried out in the laboratories of Carleton University by Dr. M. Shaffiqullah.

-
- o The error limits are quoted for the 95% confidence level.
 - oo Value missing.

APPENDIX II

EQUAL AREA PROJECTION

The plotting and contouring of the structural readings were carried out by computer (Fig. 33,34,36,38). The program employed, slightly modified by the author, was courteously made available by Dr. K.J. Rosengren of the Department of Geophysics and Geochemistry, Australian National University.

For each group of data, three different types of diagrams are displayed:

- 1) Scatter diagram
- 2) Contour diagram by Schmidt method
- 3) Contour diagram by Kamb method

The first two are in common use and need no further explanation. The Kamb method of contouring is less well known, largely because without the employment of a computer, it is time-consuming.

This method aims to estimate on a statistical basis whether or not the orientation densities are significant. The measure of statistical significance is the probability that the observed orientation density could have resulted from random sampling of a population that lacks preferred orientation. To check this probability the area A of the counter (0.01 in Schmidt method) is variable and chosen in such a way that, to quote Kamb, (1959, p. 1908), "if the population {of the readings} lacks preferred orientation, the number of points E expected to fall within a given area A , is three times the standard deviation $\{\sigma\}$ of the number of

points that will actually fall within the area $\{A\}$ under random sampling of the population. This ensures that the observed orientation densities, if obtained by random sampling of a non-preferentially oriented population will not fluctuate widely from the expected density E/A . Observed densities that differ from E/A by more than two or three times the standard deviation σ (for random orientation) are then likely to be significant.....". Therefore, the counter lines are drawn at intervals of 2 (0,2,4) where the expected density E is 3 for no preferred orientation.

The Kamb method is particularly useful for diagrams with a small number of points, because in this case the chance of spurious (statistically insignificant) orientation densities is high with the Schmidt contouring method. For diagrams with a large number of data (over 200) the Kamb and Schmidt methods do not differ appreciably.

The method is based on the assumption that the data are statistically independent, a condition rarely satisfied in structural work. The error inherent in structural measurements is also not taken into account. Thus the method does not fully eliminate statistical limitations of the data.

Explanation of the Stereonets* (Fig. 33, 34, 36, 38)

Upper Diagram: Scatter Diagram

Middle Diagram: Contour Diagram by Schmidt method. Contour intervals as listed. Shading symbols: + lowest density, 1,2,3 progressively higher densities, solid black = maximum.

* All diagrams are lower hemisphere.

Lower Diagram: Contour Diagram by Kamb method. A = counter area,
 $E = NA$ = number of points expected to fall within area A for a random
distribution, SIG. = standard deviation, Contour intervals = 0,2,4,6,...
standard deviations, Shading symbols:

1 = area with 2 SIG. density

2 = area with 4 SIG. density

3 = area with 6 SIG. density

4 = area with 8 SIG. density

" " " " " " "

Solid black = maximum.

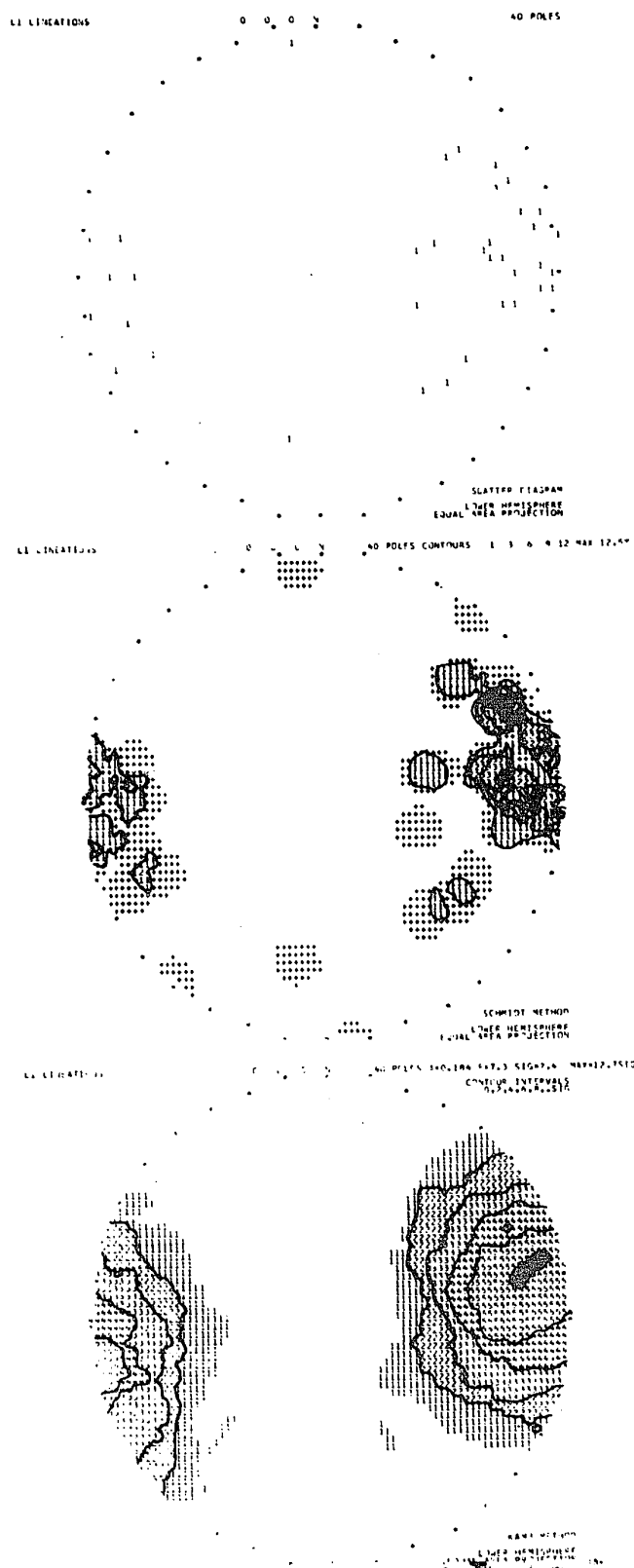


Fig. 33 Net projection of L₁ lineations in Cambro-Ordovician rocks of domain 1 (see Fig. 5). For explanation of Schmidt and Kamb contour methods see Appendix II.

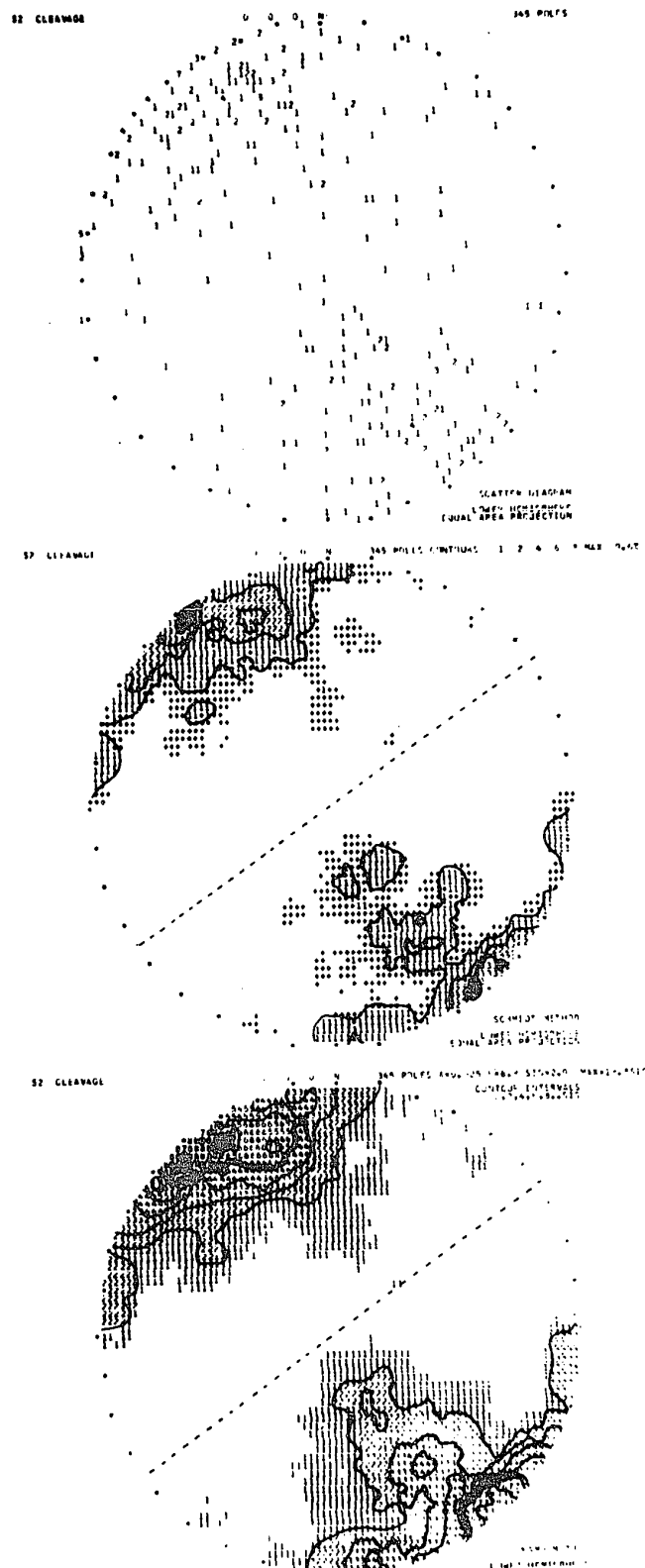


Fig. 34a Net projection of S_2 and L_2 in all Cambro-Ordovician rocks of the Mt. Albert area (domain 1 and 2 of Fig.5).
Net projection of S_2 cleavage.

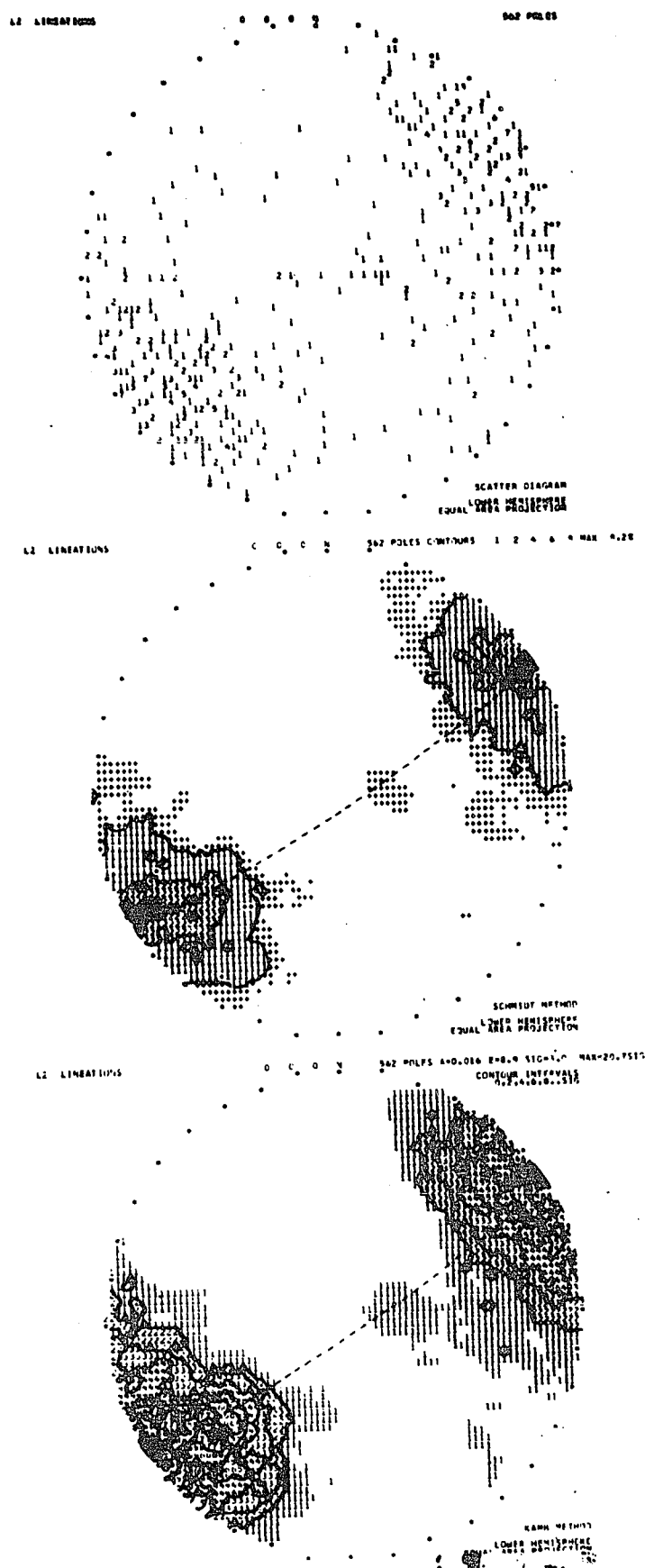


Fig. 34b Net projection of S_2 and L_2 in all Cambro-Ordovician rocks of the Mt. Albert area (domain 1 and 2 of Fig. 5).
Net projection of L_2 lineations.

Fig. 35 Map showing S_2 subareas. Subareas defined on basis of orientation of S_2 . Subarea I is characterized by S_2 cleavage subvertical, and trending northeast. Subarea II by S_2 dominantly dipping south-southeast. Subarea III by S_2 dominantly dipping north-northwest. Subarea IV by S_2 dipping east-southeast. Subarea V by S_2 consistently vertical, and trending northeast. L_2 limitations are double plunging in each subarea. Schematic S_2 stereonets in the subareas from Fig. 37 a-c. Black arrows represent modal values of L_2 lineations, (Fig. 37 f-j). Arabic numbers: map units, Roman numbers: subareas.

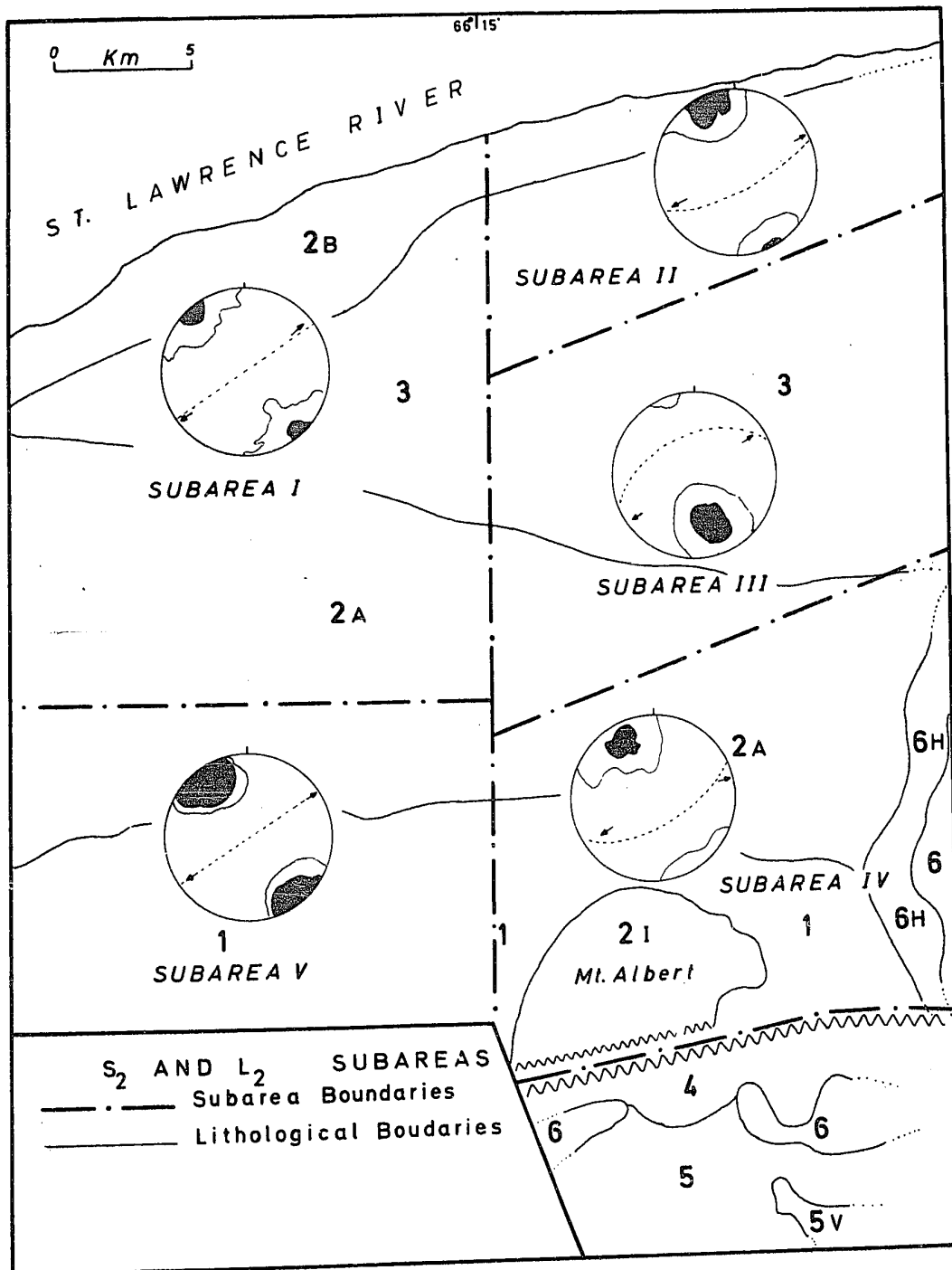


Fig. 35

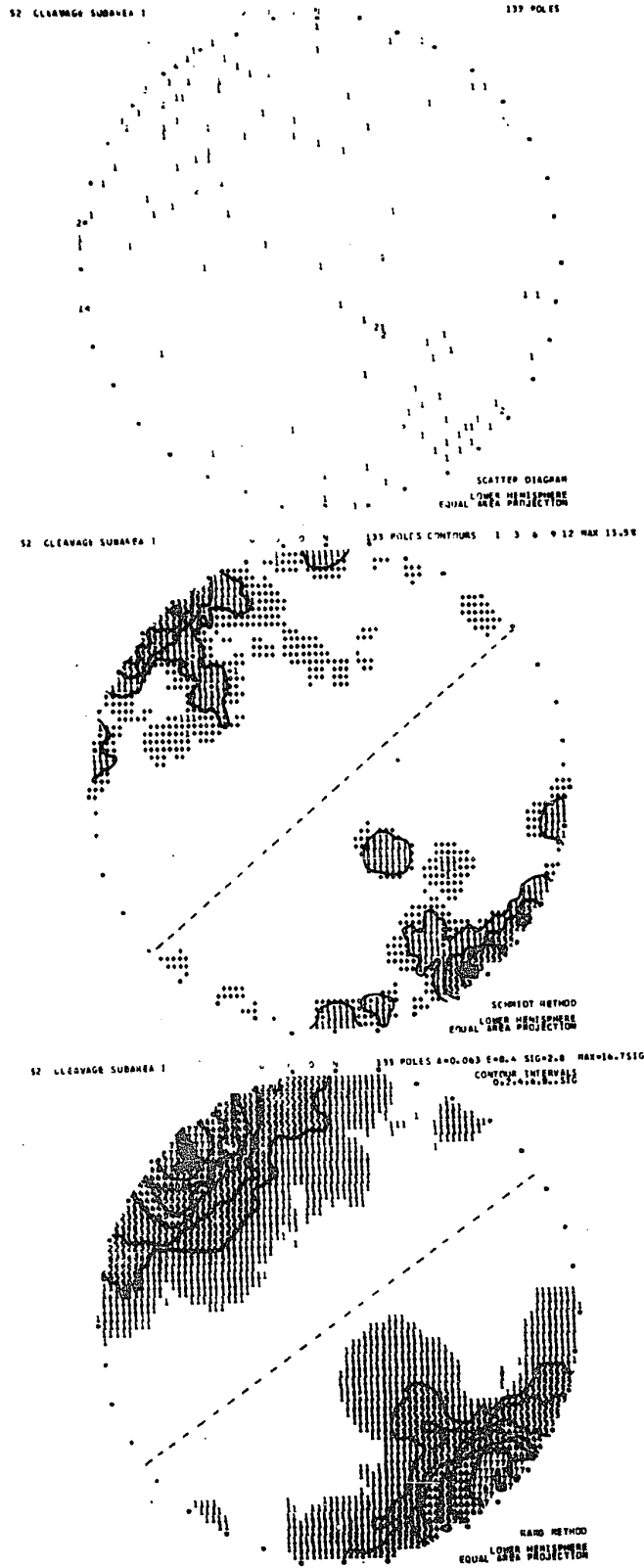


Fig. 36 a) Net projection of S_2 in subarea 1

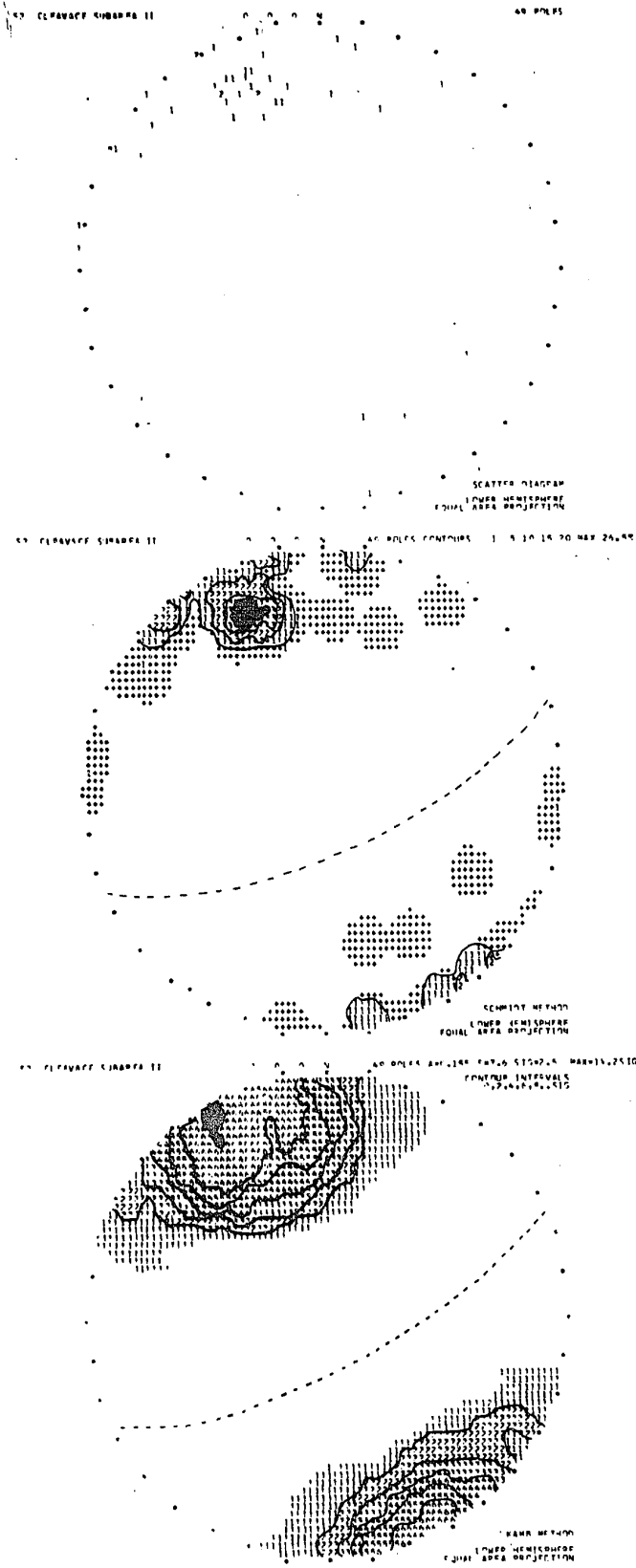


Fig. 36 b) Net projection of S_2 in subarea II

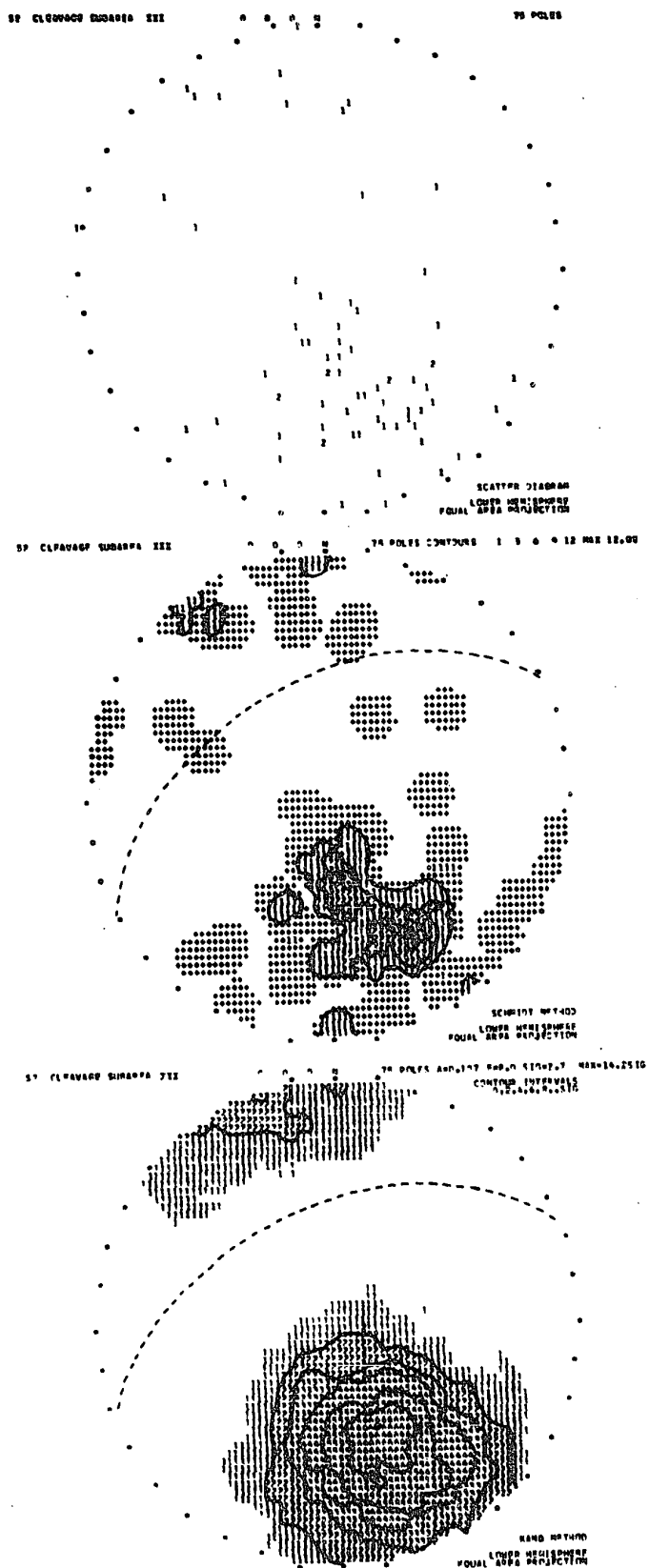


Fig. 36 c) Net projection of S_2 in subarea III

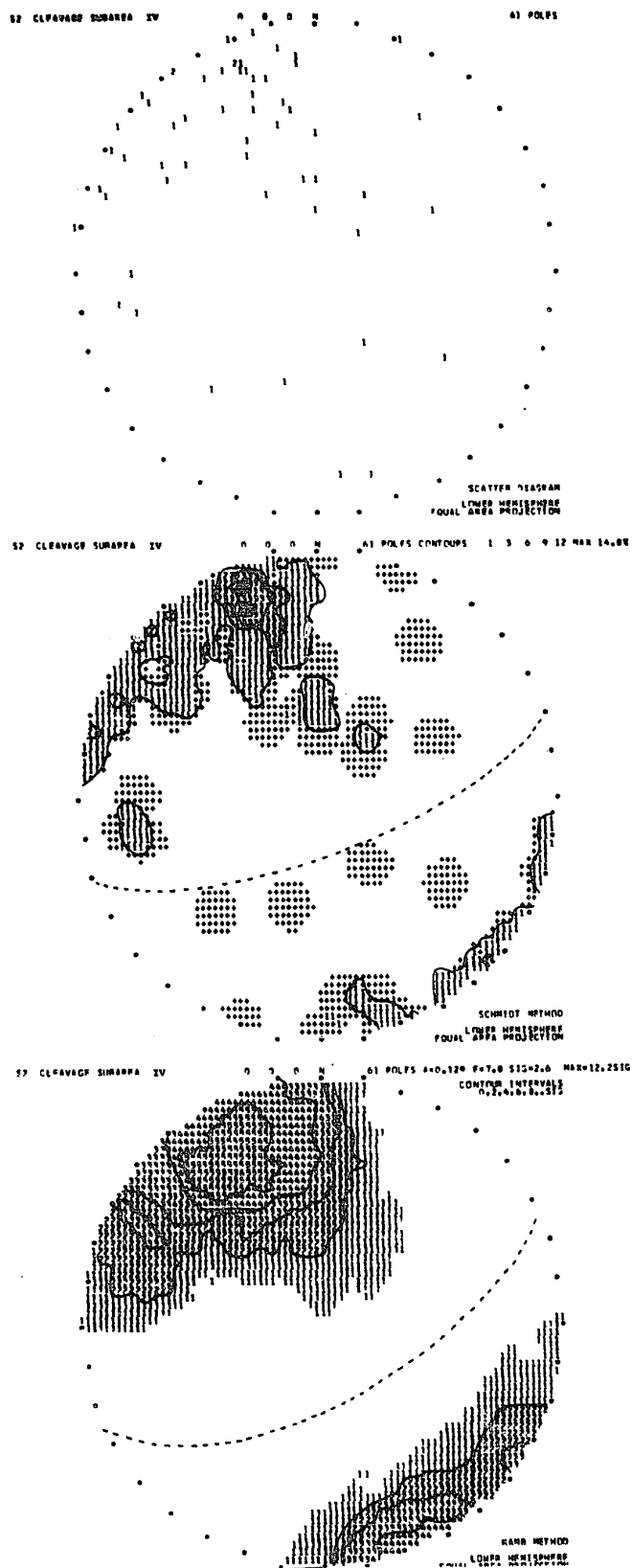


Fig. 36 d) Net projection of S_2 in subarea IV

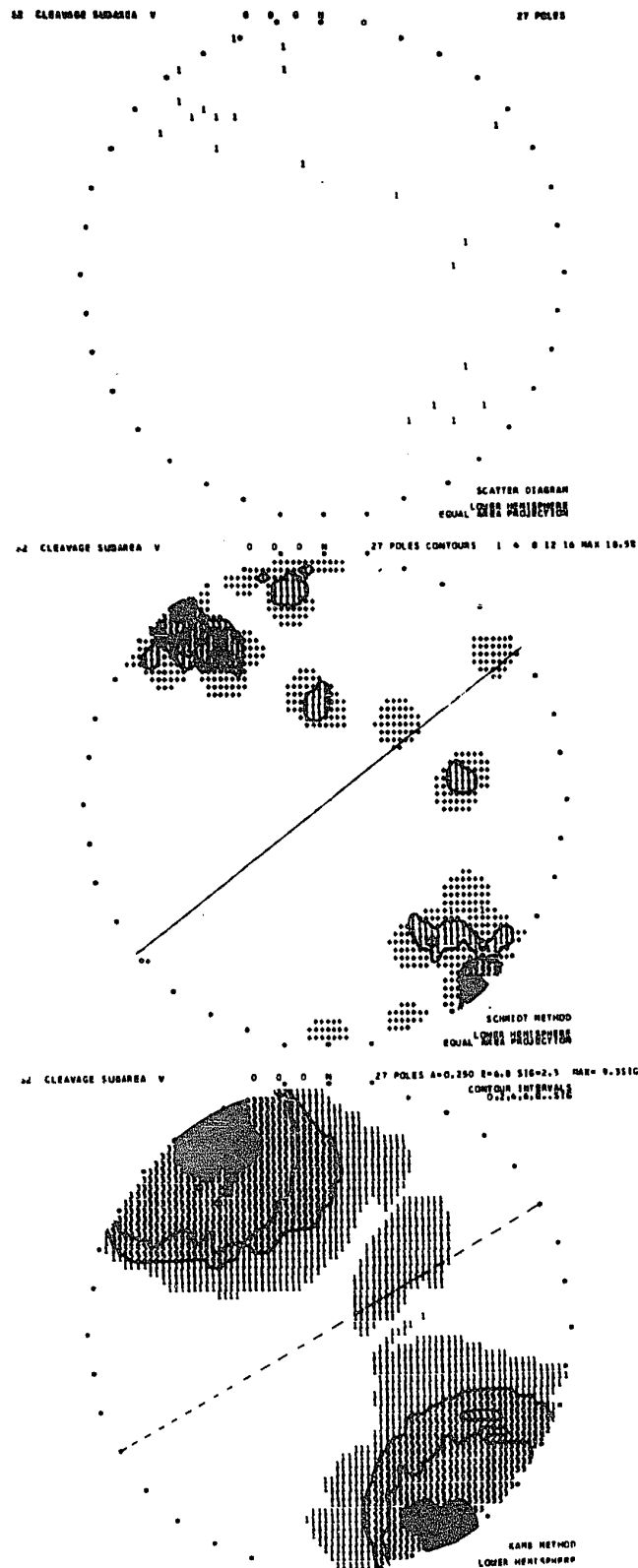


Fig. 36 e) Net projection of S_2 in subarea V

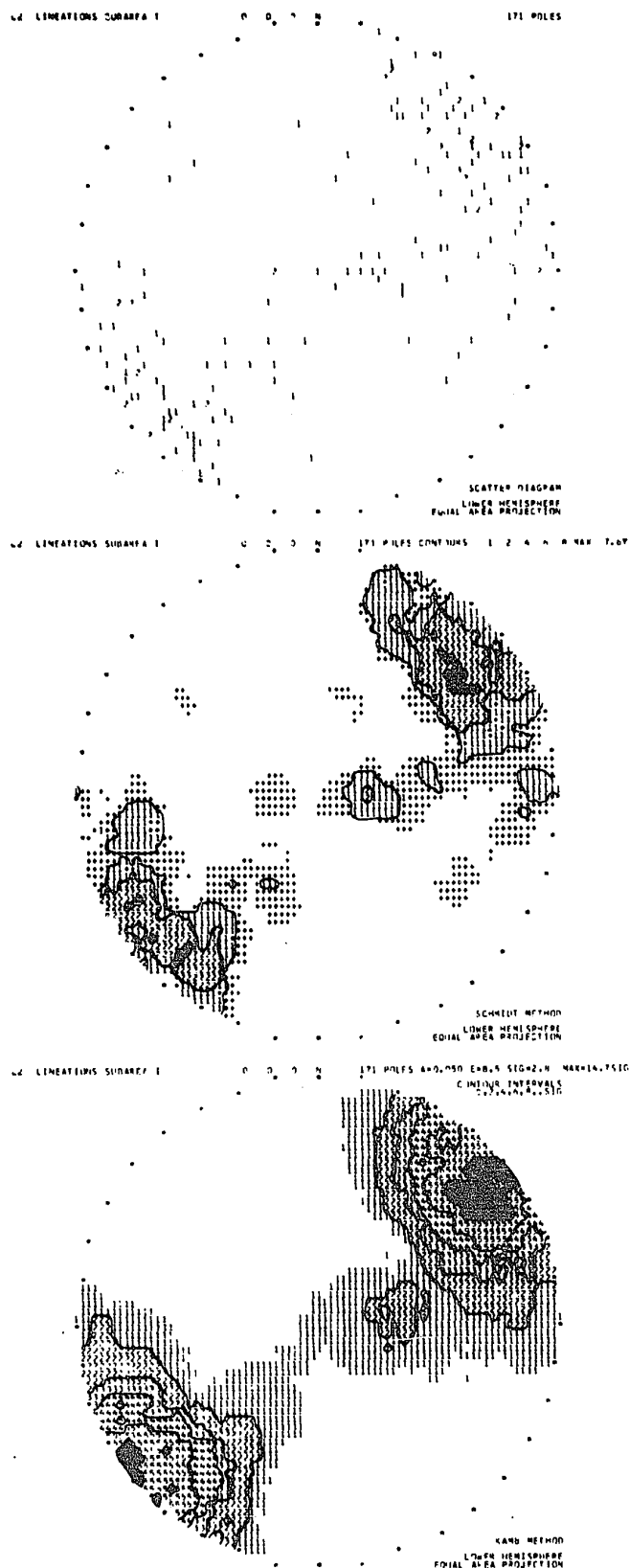


Fig. 36 f) Net projection of L_2 in subarea I

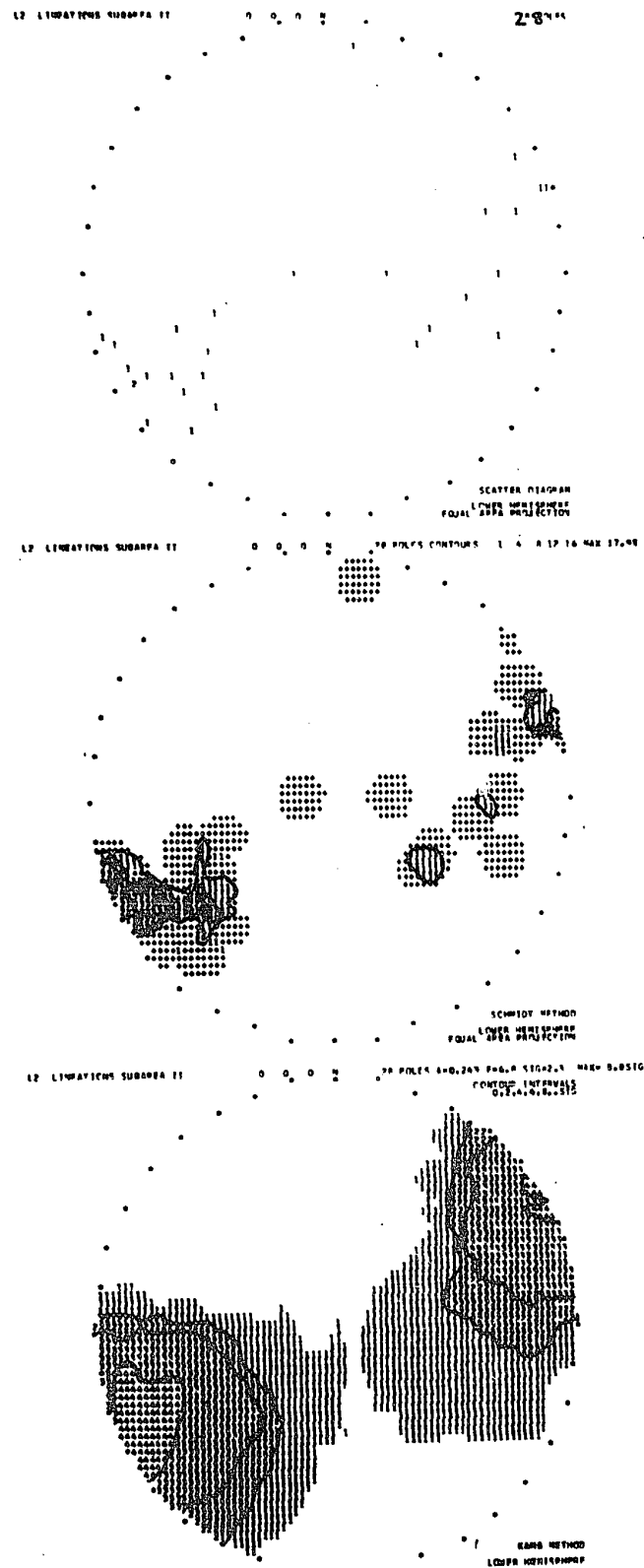


Fig. 36 g) Net projection of L_2 in subarea II

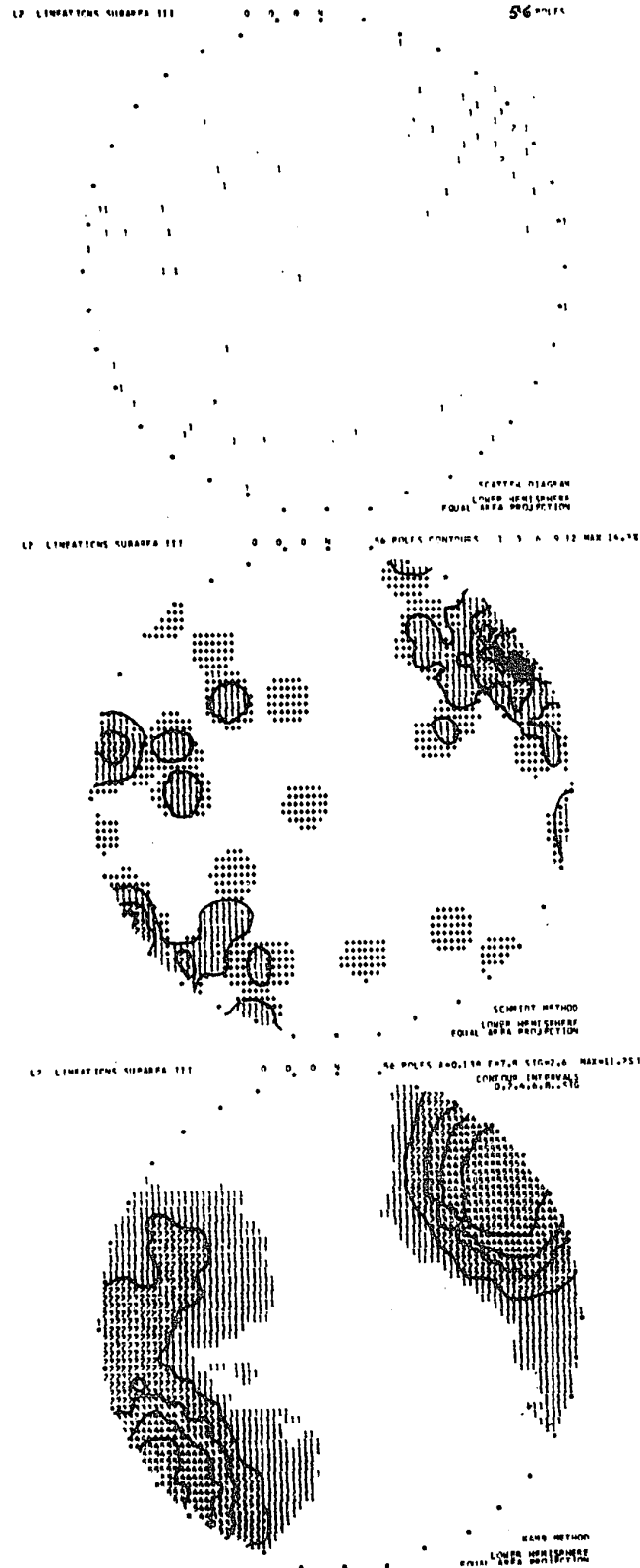


Fig. 36 h) Net projection of L_2 in subarea III

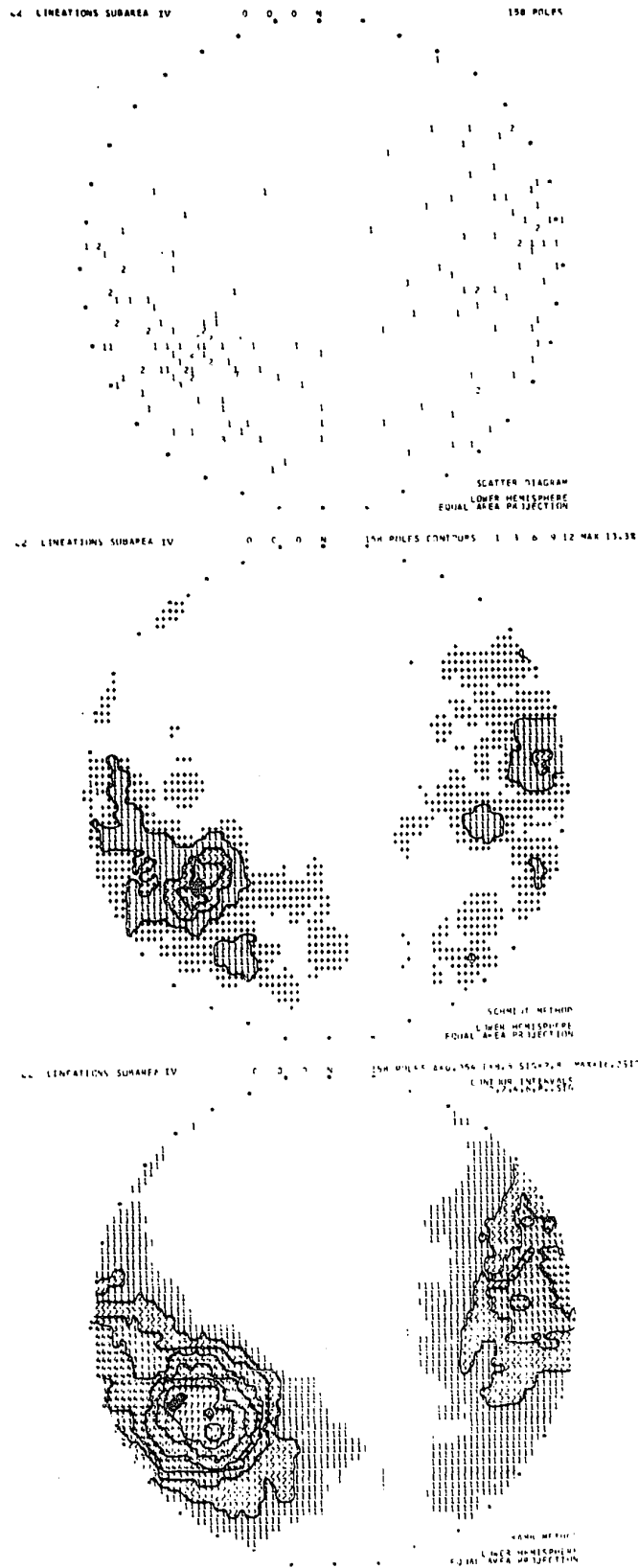


Fig. 36 i) Net projection of L_2 in subarea IV

Fig. 37 Map showing S_3 subareas. Subarea I characterized by steeply dipping discontinuous fracture cleavage, axial planar to open folds, mainly exposed along the St-Lawrence shore. Subarea II characterized by steeply dipping microkinks, kinks, and a few chevron folds, mainly present in Shickshock rocks (map unit 1). Arabic numbers: map units, Roman numbers: subareas.

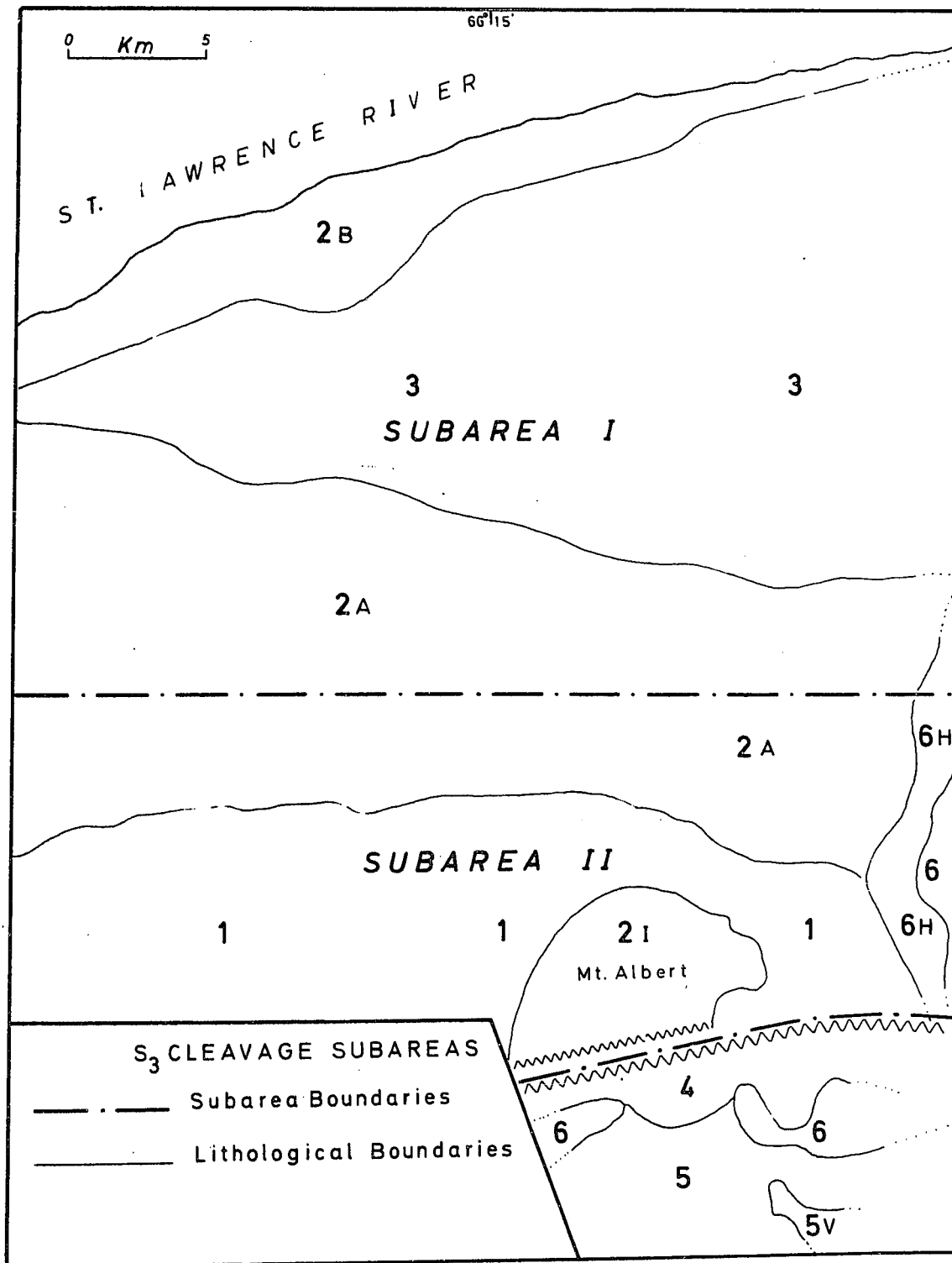


Fig. 37

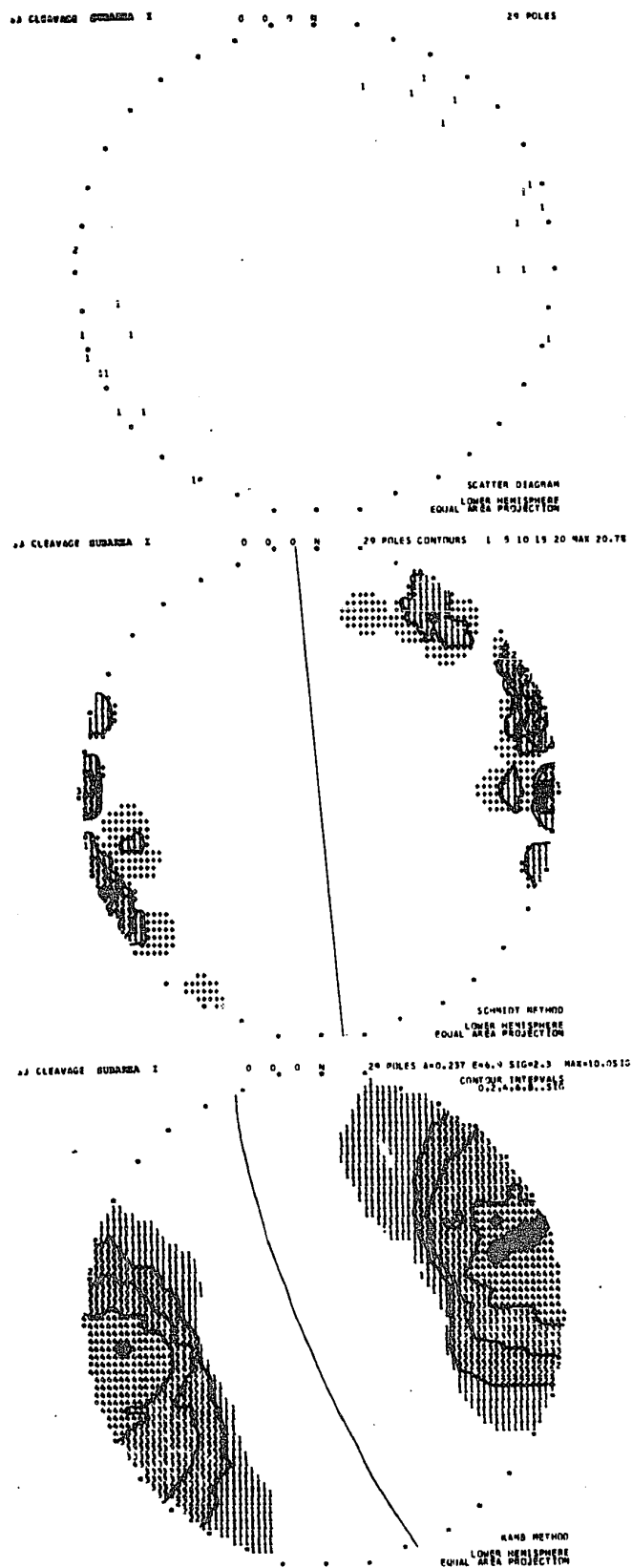


Fig. 38 a) Net projection of S_3 in subarea I

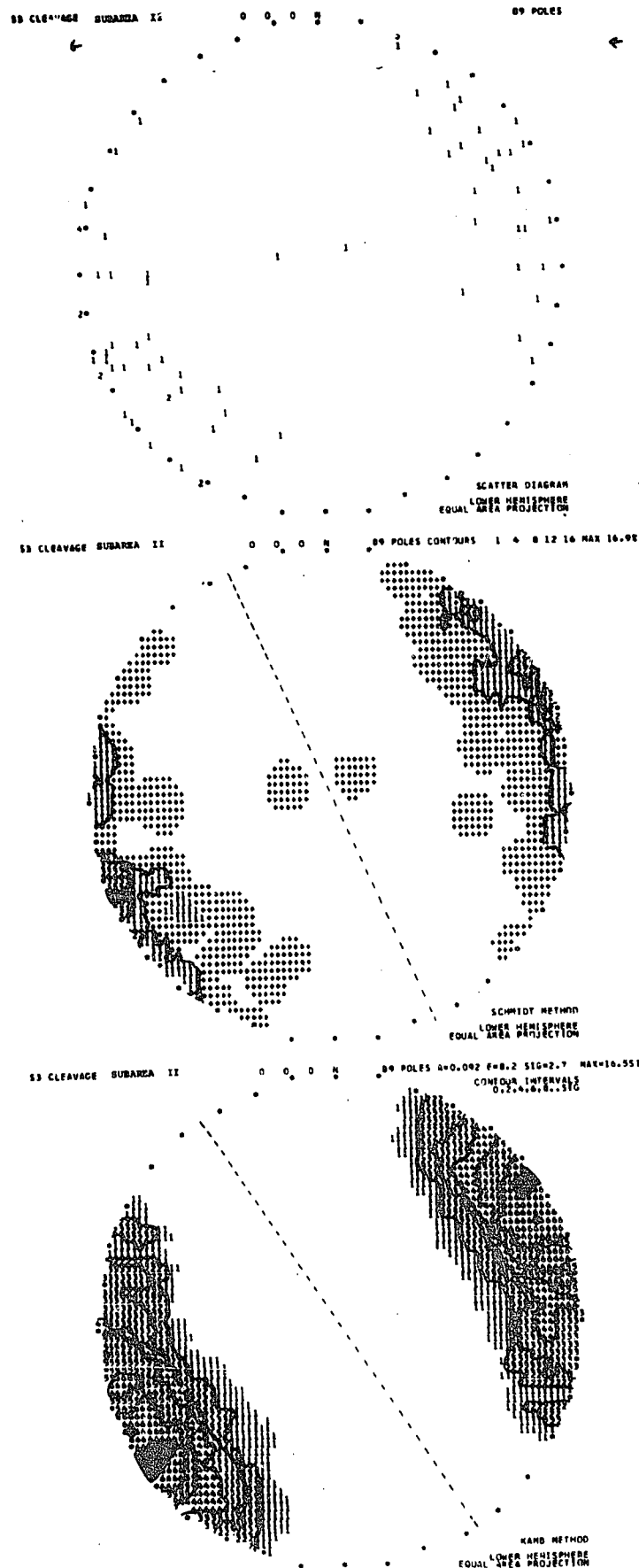


Fig. 38 b) Net projection of S_3 in subarea II

APPENDIX III

FOLD ANALYSIS

Linear Regression

A quarter wavelength was plotted against the layer thickness on log-log paper because the folds range widely in size from 1mm to 1400mm and, thus, the use of common paper was inconvenient. Because of the log-log paper used, the calculation of the regression line required a transformation to avoid the effect of the intercept of the linear equation. The regression was carried out, taking t (thickness) as an estimating variable from $W/4$ (quarter wavelength). Defining the linear regression as follows:

$$y = \beta x + \alpha$$

where

β slope of the line

α intercept the line

By using the least squares method to calculate and inserting these values in the above equation, we get

$$y = 0.579x + 13.14$$

Because of the intercept, this equation is not linear on a log-log paper.

Indeed, $\log y = \log (\beta x + \alpha)$ does not have a linear plot.

To solve this problem, a new variable z , was used where

$$z = x + \alpha/\beta$$

Therefore, the new linear equation will be: $y = \beta z$ which is the equation

of a straight line also on a log paper. Thus, first z was calculated for every x , then y was calculated for every z , and finally $\log y$ versus $\log z$ was plotted.

For example: if $x = 10$, z will be $= 10 + (13.4/0.579) = 32.69$ and y will be $0.579 \cdot 32.69 = 18.92$.

It should be pointed out that the straight line so obtained (Fig. 18) represents the best fitting for points with co-ordinates $y(t)$ and $z = (W/4) + 22.69$, and not for the actual values t and $W/4$. However, due to the log scale and to the fact that the computation of β and α is mainly dependent on the large values (100 or more) of t and $W/4$, the difference is negligible.

Correlation Coefficient

To find a measure of how "good" is the quantity $W/4$ to predict the quantity t , as standard procedure, the correlation coefficient (r) was used. Because the two populations of $W/4$ and t are strongly skewed, (see Fig. 18) to have a more reliable measurement of r , all the $W/4$ and t values were transformed to their natural logs. In this way, the two populations of $W/4$ and t deviate less from normality and r was calculated by using the transformed values.

Because the resulting r is equal to 0.96 and r for 99% of confidence level and 21 degrees of freedom is 0.516, it can be easily stated that between $W/4$ and t , there is a good correlation.

TABLE 5

Lithology, Fold Class, Interlimb Angle, and Flattening of 105 folded layers belonging to 34 folds from the Shickshock and Quebec Groups

Spec. & Layer No.		Lithology	Fold Class	Inter. Angle (Degrees)	Flattening (Percentage)
MC1	1	cal. sandstone	1B	-	0
MC3	1s	cal. sandstone	1C	45	-
"	2a	cal. sandstone	1C	35	-
34	3	cal. sandstone	1C	80	8
"	4	shale	1C	80	12
"	5	cal. sandstone	1C	80	10
"	6	shale	2	80	-
"	7	shale	3	105	-
"	8	cal. sandstone	1C	114	5
34B	3	cal. sandstone	1B	90	0
"	5	cal. sandstone	1B	110	0
"	8	cal. sandstone	1B	140	0
35	1	calcite-rich	1C	15	5
"	2	calcite-rich	1C	20	5
39B	1	calcite-rich	1C	62	10
"	2	calcite-rich	1A	55	0
51	1	quartz-rich	1C	5	5
68	1	quartz-rich	1C	5	25
"	3	quartz-rich	1C	5	5
"	4	pelite-rich	3	5	-
"	4	quartz-rich	1C	5	10
"	6	quartz-rich	1C	5	6
83C	1	quartz-calcite-rich	1C	70	15
87	2	calcite-rich	1C	55	20
"	3	calcite-rich	1C	52	15
94	1	calcite-rich	1C	30	20
"	2	calcite-rich	1C	35	15
115	1	calc. sandstone	1C	18	5
123	1	calcite-rich	1C	42	32
155	1	quartz-rich	1C	75	10
"	2	quartz-rich	3	70	-
186	1	chlor. schist	1C	24	60
"	2	chlor. schist	1C	17	40
188B	1a	quartz-rich	1C	22	12
"	2s	quartz-rich	1C	13	17
192R	1	quartz-rich	1C	82	5
"	2	sericite-rich	3	74	-
"	3	quartz-rich	1C	64	9
"	4	quartz-rich	1C	70	20
"	5	quartz-rich	1A	68	0
"	6	quartz-rich	1C	75	22

192R	7	sericite-rich	3	76	-
198	1a	chlor. schist	1C	13	45
"	2a	chlor. schist	2	13	-
"	3a	chlor. schist	3	13	-
"	1s	chlor. schist	1C	15	70
"	2s	chlor. schist	2	15	-
253	3a	calcite-rich	1C	34	33
"	7a	calcite-rich	1C	21	22
"	7s	calcite-rich	1C	23	20
253M	1a	calcite-rich	1C	21	40
"	3a	calcite-rich	1C	5	15
268C	3	calcite-rich	1C	-	12
"	5	calcite-rich	1C	-	16
"	6	pelite	3	-	-
310	1	chlor. schist	1C	40	35
"	2	chlor. schist	1C	-	-
313	1a	quartz-rich	1C	90	5
"	1s	quartz-rich	1C	80	20
330	1	cal. siltstone	1C	86	10
331	1	shale	3	35	-
"	2	calc. sandstone	1C	27	15
"	3	shale	3	25	-
386	1	quartz-rich	1C	53	10
"	6	pelite	3	50	-
400	1	calc. sandstone	1C	45	7
"	3	calc. sandstone	1C	50	30
"	4	shale	1C	54	-
400P	1	calc. sandstone	1C	42	70
409	4a	pelite	3	-	-
"	4s	pelite	3	-	-
"	6a	pelite	3	-	-
"	6s	pelite	3	-	-
"	8a	pelite	3	-	-
"	8s	pelite	3	-	-
"	9a	calcite-rich	1C	36	13
"	9s	calcite-rich	1C	35	7
"	10a	pelite	3	-	-
"	10s	pelite	3	-	-
"	12a	pelite	2	-	-
"	12s	pelite	3	-	-
"	16a	pelite	3	-	-
"	16s	pelite	3	-	-
"	17a	calcite-rich	1C	27	20
"	18a	pelite	3	-	-
"	18s	pelite	3	-	-
"	19a	calcite-rich	3	43	38
"	19s	calcite-rich	1C	49	20
"	10a	pelite	3	-	-
"	10s	pelite	3	-	-
"	21a	calcite-rich	1C	34	13
"	21s	calcite-rich	1C	51	10

450C	1a	quartzite	1C	34	35
"	1s	quartzite	1C	28	15
450	2	quartz-rich	1C	55	20
"	3	quartz-rich	1C	62	30
"	4	pelite	3	70	-
"	5	quartz-rich	1C	75	15
481F	1	quartzite	2	12	80
481	1a	quartz-rich	1C	15	60
"	1s	quartz-rich	1C	25	55
MC2	1	calcite-rich	1C	49	5
"	2	calcite-rich	1C	51	25
"	3	calcite-rich	1C	54	20
"	4	calcite-rich	1C	55	23

TABLE 6

Average Interlimb Angle and Flattening for 7 F_1 and 27 F_2 folds from the Shickshock and Quebec Groups

Specimen	Intel. Angle	Flattening %	Fold Gener.
MC1	-	0	F_2
MC3	40	-	F_2
34	89	-	F_2
34B	113	0	F_2
35	17	5	F_2
39B	58	5	F_1
51	5	50	F_2
68	5	11	F_1
83C	70	15	F_2
87	53	17	F_2
94	32	17	F_2
115	18	5	F_2
123	42	32	F_2
155	72	10	F_2
186	20	50	F_2
188B	17	17	F_2
192R	73	22	F_2
198	13	57	F_1
253	26	25	F_1
253M	13	27	F_1
268C	-	14	F_2
310	40	35	F_2
313	85	13	F_2
330	86	10	F_2
331	27	15	F_2
386	53	10	F_2
400	50	12	F_2
400P	42	70	F_2
409	39	17	F_1
450C	31	25	F_2
450	65	18	F_2
481F	12	80	F_2
481	20	57	F_1
MC2	52	18	F_2

APPENDIX IV

PRELIMINARY STRUCTURAL INVESTIGATION IN TEMISCOUATA AREA

Introduction

In the summer of 1971, a preliminary structural survey was carried out in the region, about 250 kilometers southwest of the Mt. Albert area, that approximately lies between the towns of Rimouski, Rivière du Loup, and Cabano, which after the well-known lake, is here referred to as the Témiscouata area (Fig. 3g).

The aims of this investigation were:

- 1) To investigate whether the structural history of Paleozoic rocks in the western part of Gaspé Peninsula, is similar to that established in northeastern Gaspé. The resultant information is pertinent to the general problem of correlating small-scale structures between distant areas.
- 2) As a part of 1., to find if the early isoclinal recumbent folds (F_1 Taconic structures), recognized northeastward in Matane (Sikander, 1967) and in the Mt. Albert areas, appear also in this area. This fact is considered of interest because to the east the early folds are associated with a belt of metamorphic rocks (Shickshock Mountains) and ultramafic intrusions, whereas in the Témiscouata area, there are no such intrusions and the rocks are at a very low grade of regional metamorphism.
- 3) Analysis of the dominant (F_2 -type) folds in rocks of the Quebec Group (Cambro-Ordovician) and in adjacent Siluro-Devonian terrains, deformed respectively in the Taconic and Acadian orogenies in order to

compare within a restricted area and in similar lithologies, the style of folds of different orogenic events.

4) To determine the effects of the Acadian orogeny on rocks already deformed during the Taconic revolution.

Stratigraphy

The complex stratigraphy of the study area has been defined by several workers (Béland, 1960, Lajoie, 1961, 1962, Lespérance and Greiner, 1969). A stratigraphic and paleogeographic synthesis of the Silurian formations was carried out by Lajoie, et al, (1968).

Quebec Group

The supposed older (Cambrian?) rocks of the Quebec Group are composed mainly of gray, red, phyllitic slates and siltstones, with green quartzites and impure locally predominant sandstones. The rocks display a foliation, generally parallel to the bedding, defined by the alignment of chlorite, muscovite, and strained quartz grains.

The upper part of the Quebec Group (Trinité Group, Lajoie, 1961, 1962) of Lower Middle Ordovician age is formed of limestone conglomerate, gray impure sandstones, red, green, gray argillites, and gray quartzites.

Silurian Formations

A thick sequence (from 2000 to 6000m) of Silurian detrital rocks rests unconformably on the Quebec Group. The contact between the Silurian and older rocks is generally marked by faults, however, at least at

two localities (near St. Guy, 40 kilometers north of Cabano, and near St. Mathieu, midway between Rimouski and Rivière du Loup) an angular unconformity is ascertained.

According to Lajoie et al., (1969), the early Silurian sea transgressed from northeast to southwest over and around the partly eroded "Taconic Mountains". The deposition of terrigenous coarse to fine grained gray conglomerate, sandstone and siltstones (Cabano Formation), was followed by deposition of clastic materials of volcanic origin, (Pointe Aux Trambles-Lac Raymond Formations).

However, the major transgression of the Silurian began later, during the late Middle Silurian, with the deposition of red and gray quartzites and limestone conglomerate (Robitaille Formation).

Devonian Formations

By comparison with the underlying Silurian formations, the Devonian rocks are much more pelitic. Argillaceous limestone, shale, shaly siltstones with minor limestone, dolomite, and quartzite (Témiscouata Formation, Cap Bon Ami Formation) of Lower Devonian age form a thick sequence (about 4000 m) which outcrops mainly in the southern part of the study area. These rocks lie along the axis of the Gaspé-Connecticut synclinorium.

It has been suggested that a disconformity marks the boundary between the Silurian and the Devonian (Lespérance and Greiner, 1969).

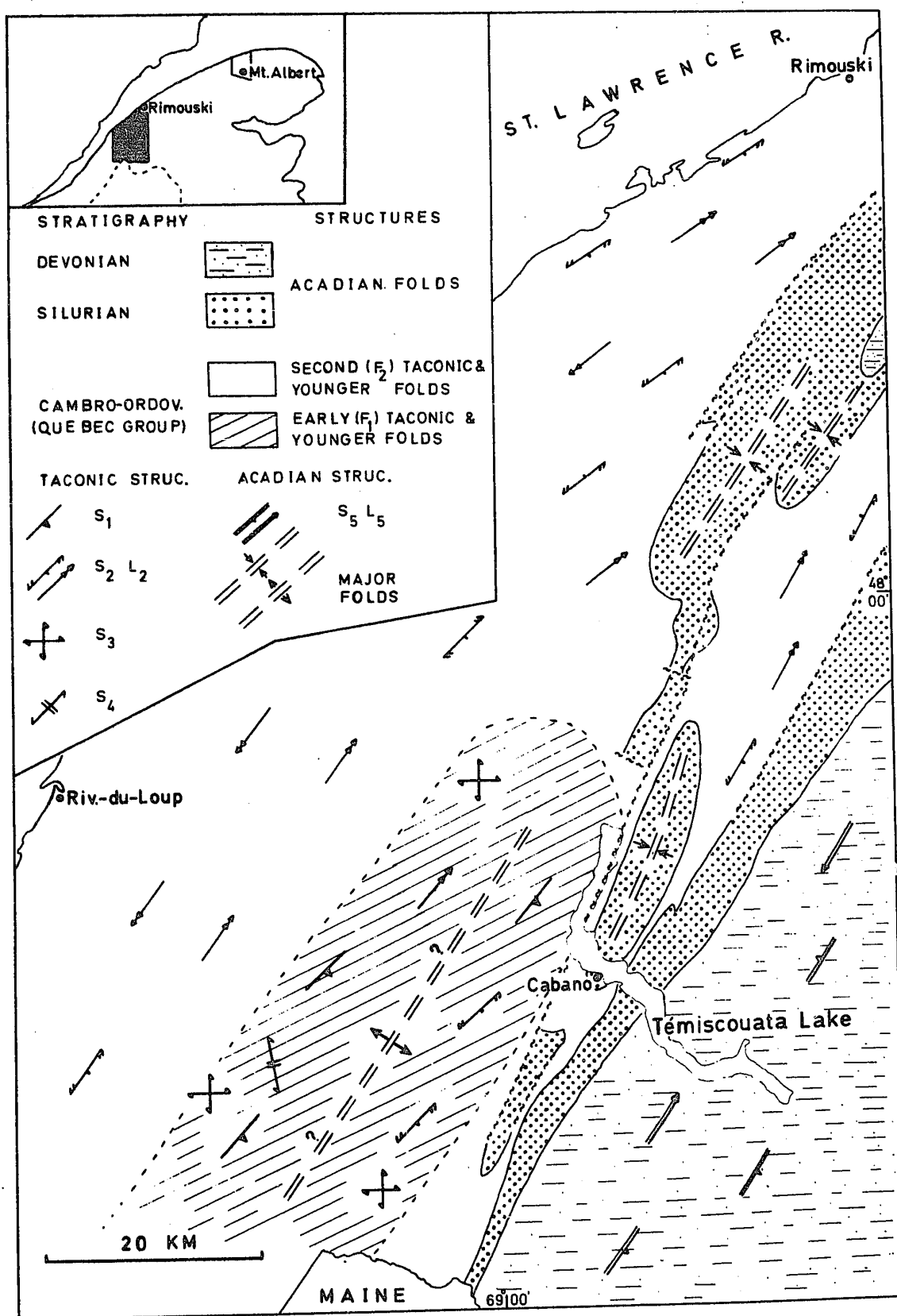


Fig. 39 Geologic Map of Témiscouata Area. Modified after Lajoie (1961, 1962), and Lespérance and Greiner (1969).

Structural Geology

Structures of five deformations of varying character were recognized in the investigated area. All but the last, are considered as Taconic features because they are confined to the Cambro-Ordovician.

An early tectonic event formed an S_1 slaty cleavage and schistosity that are axial planar to isoclinal recumbent (?) folds (F_1). The development of the schistosity in pelitic quartzites and of the slaty cleavage sensu stricto in pelites, appears to have been enhanced by areally discontinuous low grade dynamo-metamorphism.

The trend of F_1 is not yet clearly known, because too few axial lineations and folds were detected. However, it seems that before later deformation, these folds were shallow plunging and northeasterly trending.

The areal distribution of these folds has been approximately delineated (Fig. 39). It was found difficult and sometimes impossible, to separate the early foliation (S_1) from a bedding fissility and other younger planar fabrics. Nevertheless, on the basis of a few exposures where evidence was unequivocally supported by overprinting relationships, it appears that the rocks affected by these early folds form a northeast trending belt about 20 km. wide. This belt could extend eastward under the adjacent Silurian formations; however, it could also be truncated by a fault.

As in Mt. Albert area, the contact between rocks affected and unaffected by F_1 is not marked by an apparent unconformity. Moreover, the boundary separates rocks of identical lithology.

The second folds affect all the pre-Silurian rocks of the area, (Fig. 39). These folds, varying in wavelength from a few cm to 2-3 Km, are generally tight, upright, or slightly overturned to the northwest and doubly plunging. Moving from the northeast to southwest of the area, the axes of these structures swing from N 50° - 60° E to N 30° - 40° E.

Similar folds can readily be traced northeast along the shore of the St. Lawrence River to the Mt. Albert area, where they are also F_2 structures.

Fracture cleavage and slaty cleavage sensu lato, are associated with F_2 folds where S_1 is absent; where S_1 is the folded surface, a discontinuous crenulation cleavage was observed. It, thus, appears that S_2 cleavage developed better (at least macroscopically) in rocks unaffected by F_1 .

Structures of the third deformation are discontinuous throughout the area, but locally, penetrative at the outcrop scale. Most prominent is a shallow or horizontal crenulation cleavage that is axial plane to open or gently recumbent folds and to conjugate (?) sets of kink-bands. These structures, almost restricted to the F_1 belt (Fig. 39), are shown to be younger than the F_2 folds by the fact that the horizontal cleavage crenulates S_2 as well as S_1 surfaces.

Structures of a possible fourth deformational event are present in few localities. Most notable is a steeply dipping crenulation cleavage affecting S_1 , S_2 and S_3 , and striking northwest. No associated larger folds have been recognized.

Structures attributed to the last deformational event of Acadian

age are most obvious in both the Silurian rocks, which unconformably overlie the Cambro-Ordovician rocks tightly folded by F_2 , and the Devonian rocks. The Silurian sandstones form open to gentle, upright, northeasterly trending folds characterized by an imperfect fracture cleavage (Plate 10a). In the pelitic overlying Devonian rocks, folds of similar orientation are open to tight, and there is a well-developed, penetrative fracture cleavage and slaty cleavage sensu lato, which is similar to that in F_2 folds, (Plate 10b).

Large-scale northeast trending folds with wavelengths of several kilometers affect the outcrop pattern, and it is possible that in the Cambro-Ordovician rocks, the belt of F_1 structures, which has a similar trend, represents a contemporaneous Acadian anticlinal feature. (Fig. 39). However, the earlier F_2 folds are also of this trend, so that from outcrop patterns it is difficult to distinguish structures of different ages.

Conclusions

On the basis of these preliminary observations, the following conclusions can be drawn:

Within the Cambro-Ordovician rocks there is a 20 km wide, northeast trending belt in which an early schistosity and slaty cleavage lie parallel to bedding in the same manner as S_1 in the Mt. Albert area. Probably the foliation also developed in an early stage of the Taconic orogeny. The boundary of the belt does not seem to be marked by an unconformity. It is suggested that the early structures were "basement" features brought up by gentle large-scale warping, of probably Acadian

age, which accounts for the present outcrop pattern.

Taconic (F_2) and Acadian (F_5) folds in rocks of comparable lithology, are similar in shape, attitude, and display similar cleavage. Without stratigraphic evidence, they could be easily ascribed to the same deformational event.

Because the shallow cleavage of the third deformation appears to be developed on older steeply dipping foliation, it is suggested that this feature could be caused by tangential stress release and subsequent vertical compression by gravity. (Similar hypotheses have been suggested by other authors (Ramberg and Ghosh, 1968), to account for late flat foliations.

The effect of the Acadian orogeny on the pre-Silurian rocks may be restricted to a gentle warping of the Quebec Group sedimentary rocks, which formed the anticlinorium of F_1 structures.

It is notable that the cleavage of the main Acadian folds was not observed superimposed on the folded rocks of the Quebec Group. The Taconic belt possibly behaved as a "rigid basement" during the mid-Devonian tectonism. The several major faults of post-Ordovician age, which occur throughout the area may support this hypothesis.

Plate 1 Structures and textures in the Shickshock metavolcanics.

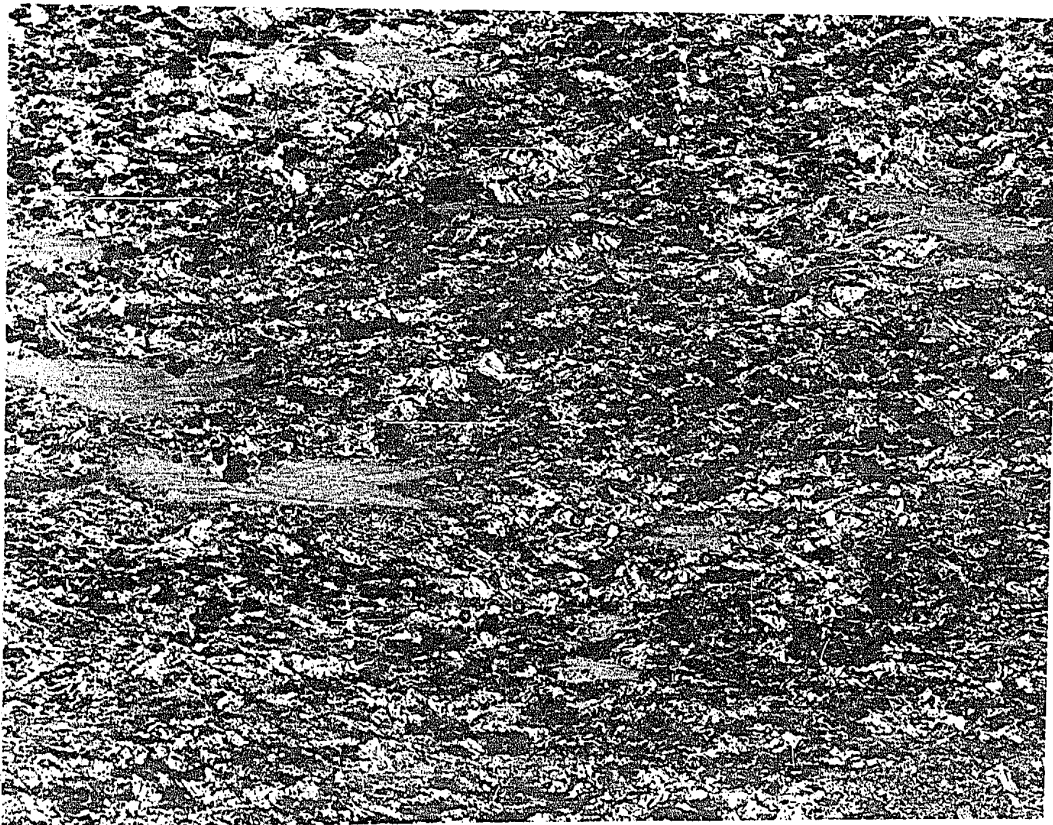
- a) Boudinage of hornblende-rich layers along S_1 foliation in feldspar-chlorite-epidote-quartz schists, (loc. 186).

- b) L_1 lineation defined by elongated chlorite porphyroblasts in a medium grained chlorite-albite-epidote-quartz matrix (x 10), (loc. 453).

PLATE 1



a)



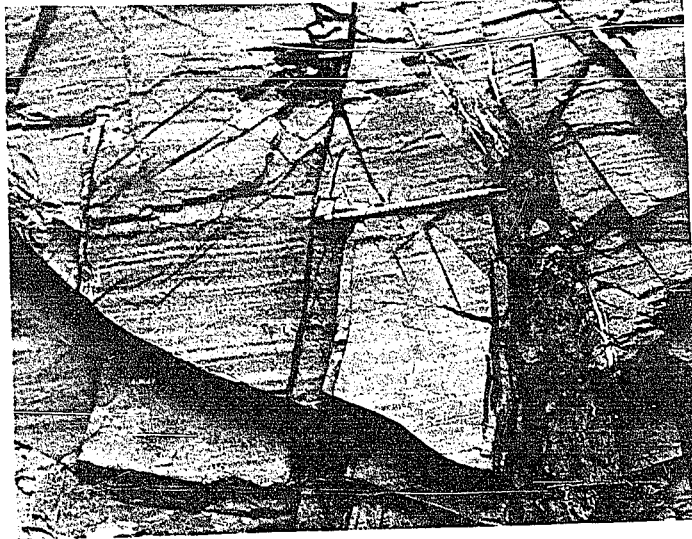
b)

Plate 2 D_1 and D_2 structural features.

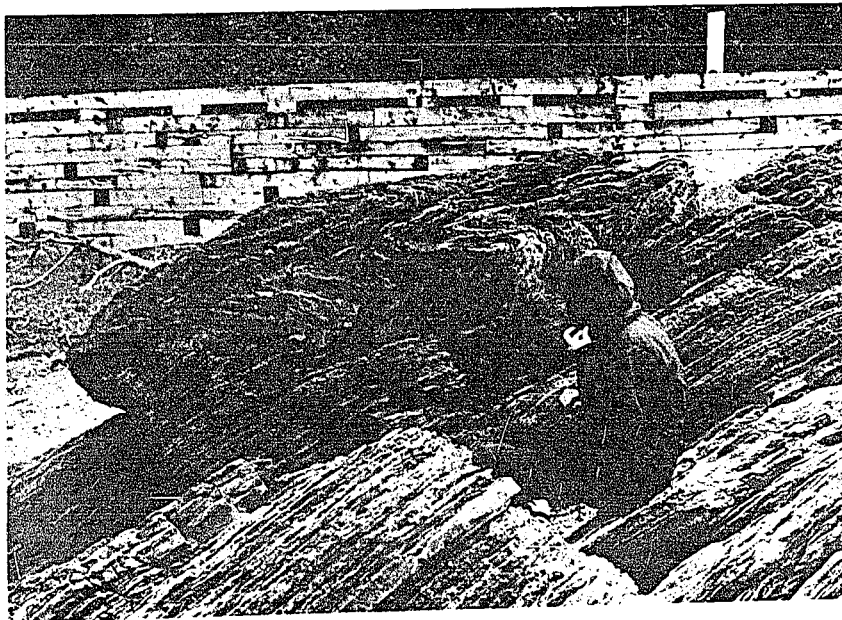
- a) Bedding (S_0) defining a series of dark and light bands (L_1) on cleavage (S_1) surface. Phyllitic-slate unit (loc. 211).

- b) Small-scale F_2 folds in calcareous siltstones interbedded with red argillites of the argillite-greywacke unit (loc. 120). Folds overturned northwest (to right hand side of photo).

PLATE 2



a)



b)

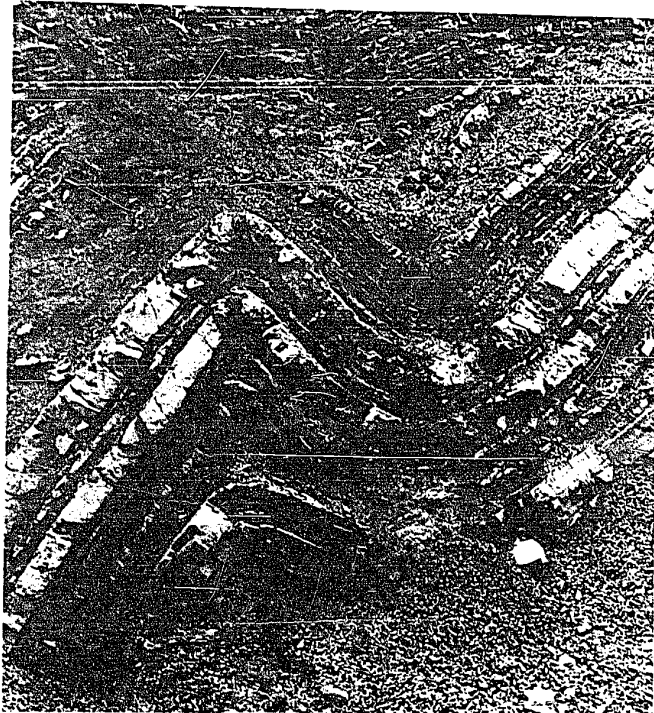
Plate 3 F_2 folds in coastal exposures of the argillite-greywacke unit.

- a) Upright open anticline and syncline with angular and rounded hinges, trending north-northeast, in sandstones interbedded with gray shales (hammer in the core of the anticline), (loc. 370).



- b) Disharmonic folding in sandstones, calcareous sandstones in a "shaly matrix". The axial planes of the folds, steeply dipping southeast (to the left) in the lower part of the outcrop, gradually flatten upward. Polyclinal folding in the top left of the exposure (loc. 370).

PLATE 3



a)



b)

Plate 4 Small-scale F_2 disharmonic folding in the argillite-greywacke unit.

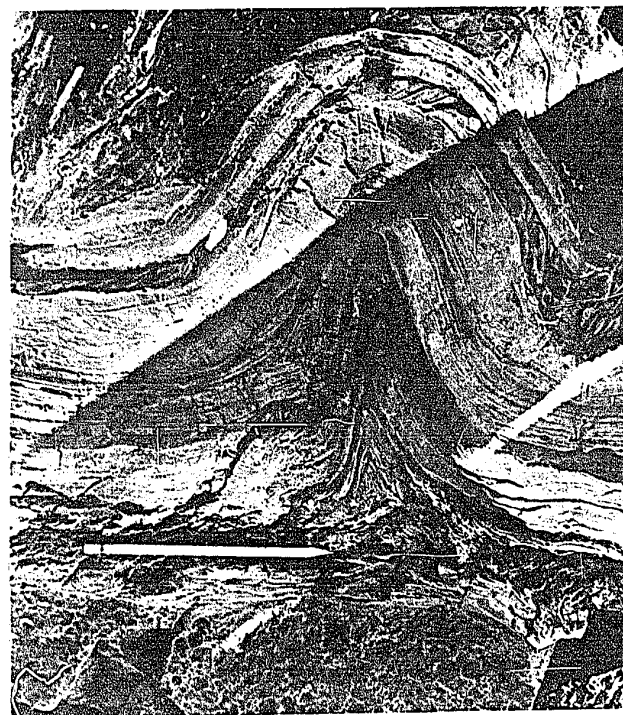
a) Disharmonically folded laminated calcareous sandstones interbedded with thin shaly layers (loc. 118).

b) Anticline in fine-grained sandstone dying downward and lying on an unfolded limestone conglomerate bed. The interbedded shales acted as a décollement surface (loc. 118).

PLATE 4



a)



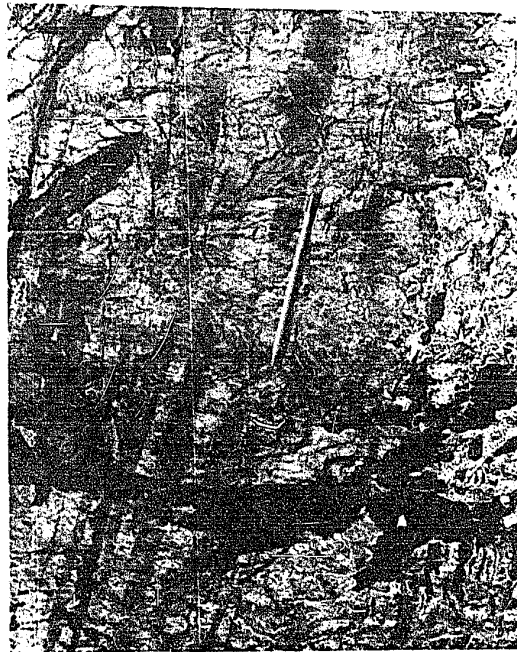
b)

Plate 5 D_3 deformation and D_2 segregation features.

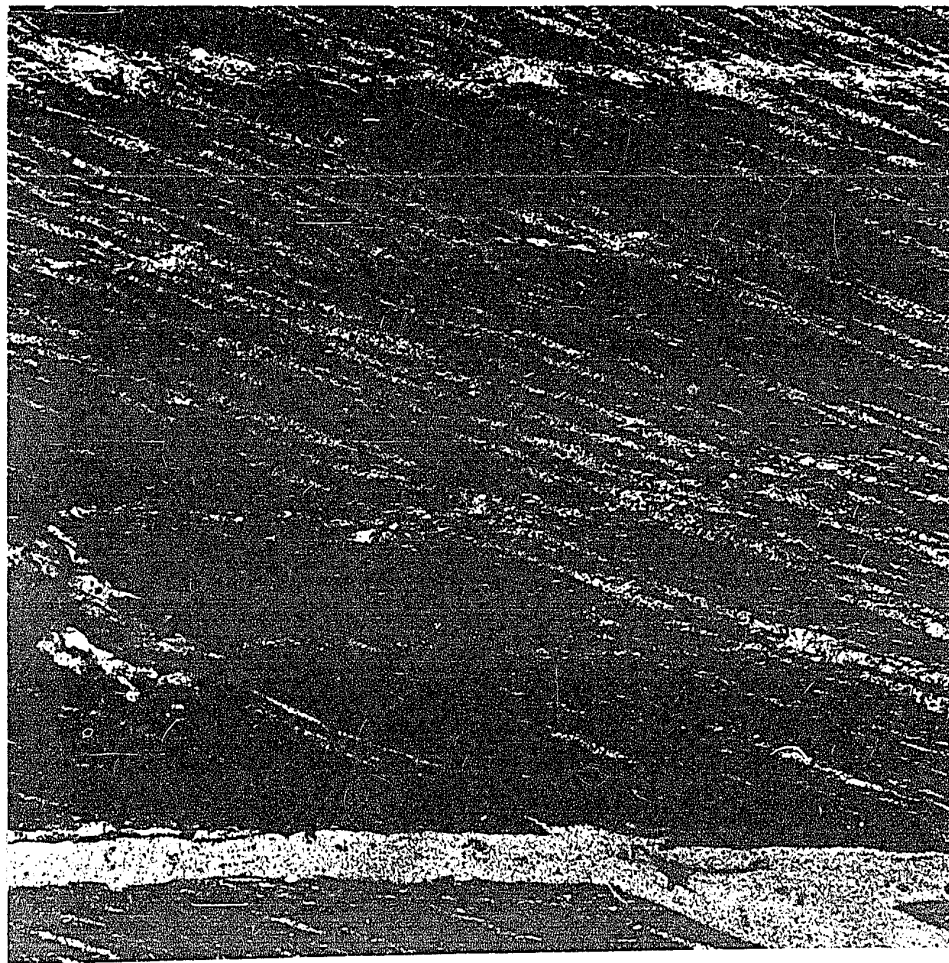
a) Steeply dipping F_3 kink bands in phyllites of the phyllitic-slate unit (loc. 3405).

b) Fine quartz segregation bands along S_2 crenulation cleavage. Phyllitic-slate unit. Quartz veins along S_1 (horizontal) and S_2 (x 25), (loc. 430).

PLATE 5



a)



b)



Plate 6 Garnet crystals in metasediments within Mt. Albert contact aureole. Garnet with internal fabric (S_1) of aligned quartz inclusions. Matrix grains of quartz and muscovite form S_1 foliation. S_1 in garnet at high to low angles to external S_1 in the matrix suggesting variable amount of rotation of garnets (x 10), (loc. 516).

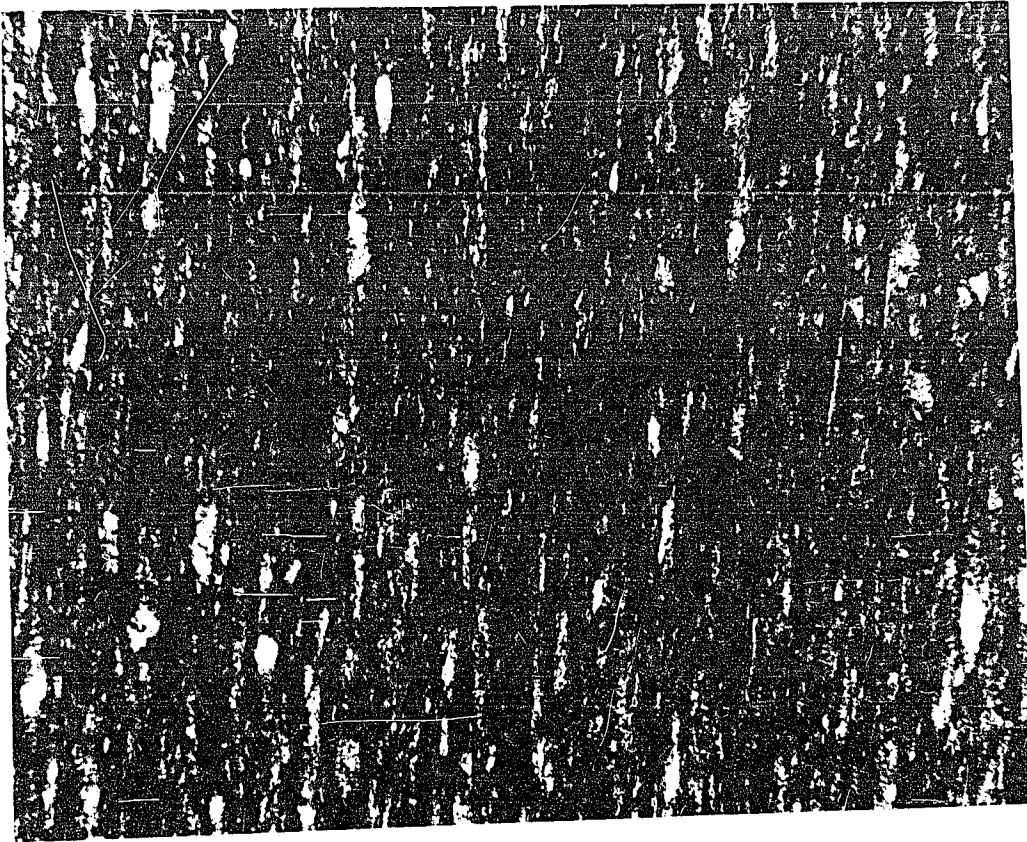
PLATE 6



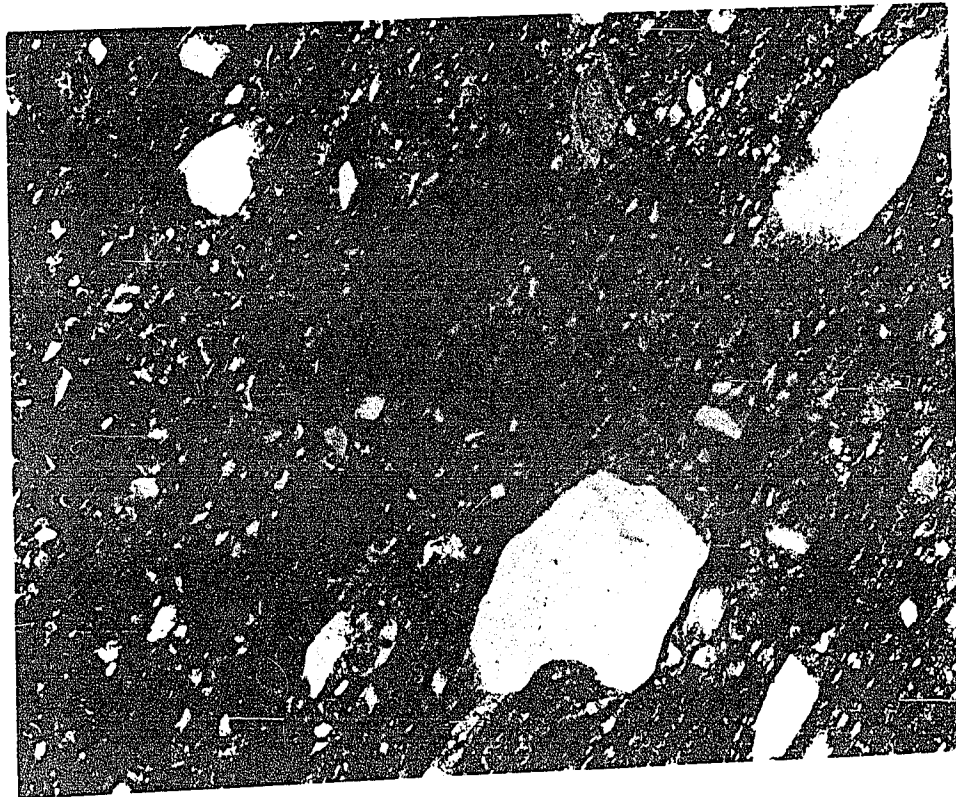
Plate 7 Cleavage types in sedimentary and low-grade pelites and semi-pelites.

- a) Slaty cleavage sensu stricto (S_1). Phyllosilicates and quartz grains (lensoidal) are homogeneously oriented throughout the investigated area (x 100). Phyllitic-slate unit (loc. 70).
- b) Slaty cleavage sensu lato (S_2) defined by the alignment of sericite and chlorite in narrow domains (shown as dark lines) in sandy shales of the siltstone unit. Quartz grains have tails containing less well oriented flaky minerals (x 60) (loc. 340).

PLATE 7



a)

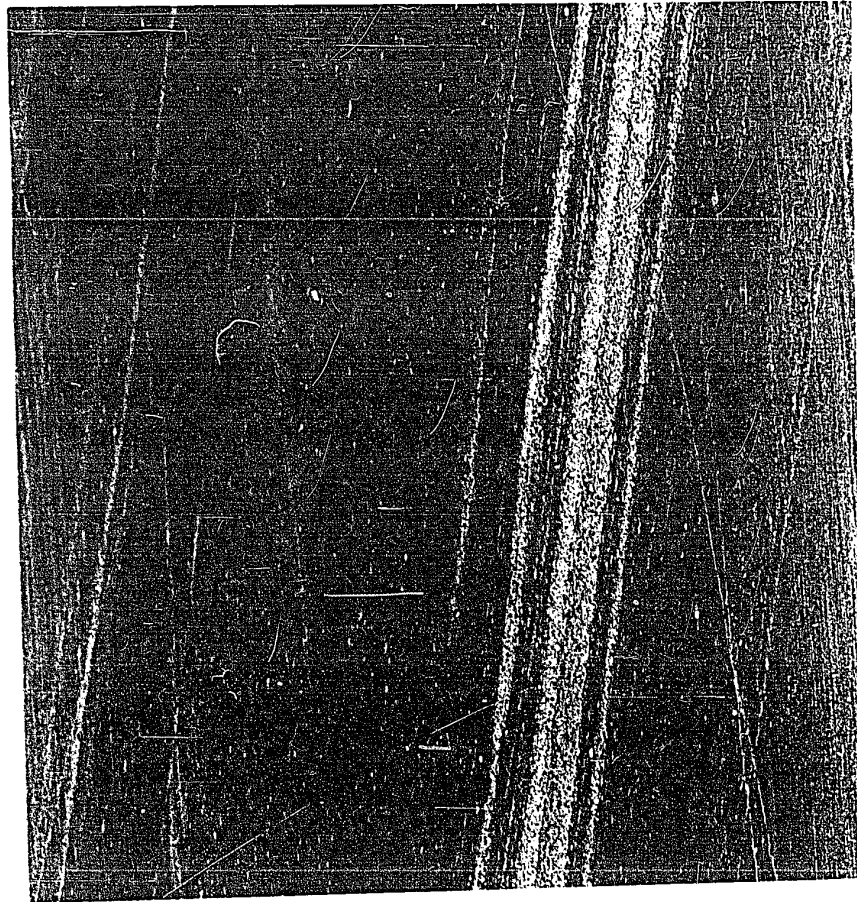


b)

Plate 8 Cleavage types in low-grades pelites.

- a) Slaty cleavage sensu stricto (S_1) defined by well aligned chlorite and sericite at small angle to the bedding (S_0) shown as light and dark bands of variable width (x 5). Phyllitic-slate unit, (loc. 145).
- b) Calcite segregation laminae (horizontal) parallel to slaty cleavage sensu stricto (S_1). Thin calcite-rich beds (S_0) are partially disrupted and transposed parallel to S_1 (x 10). Phyllitic-slate unit, (loc. 433).

PLATE 8



a)



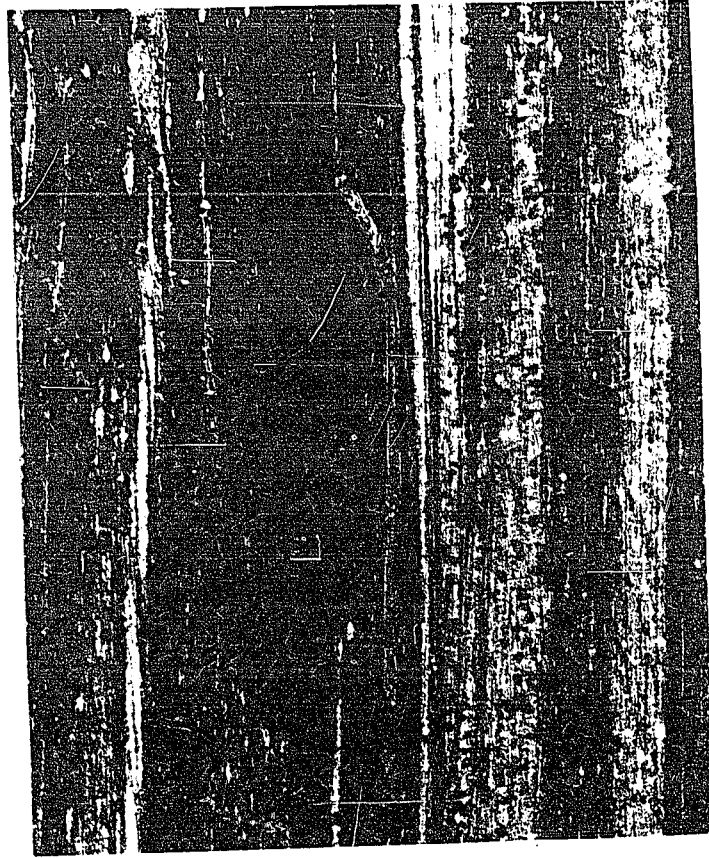
b)

Plate 9 Cleavage types in sedimentary and low-grade pelites and semi-pelites.

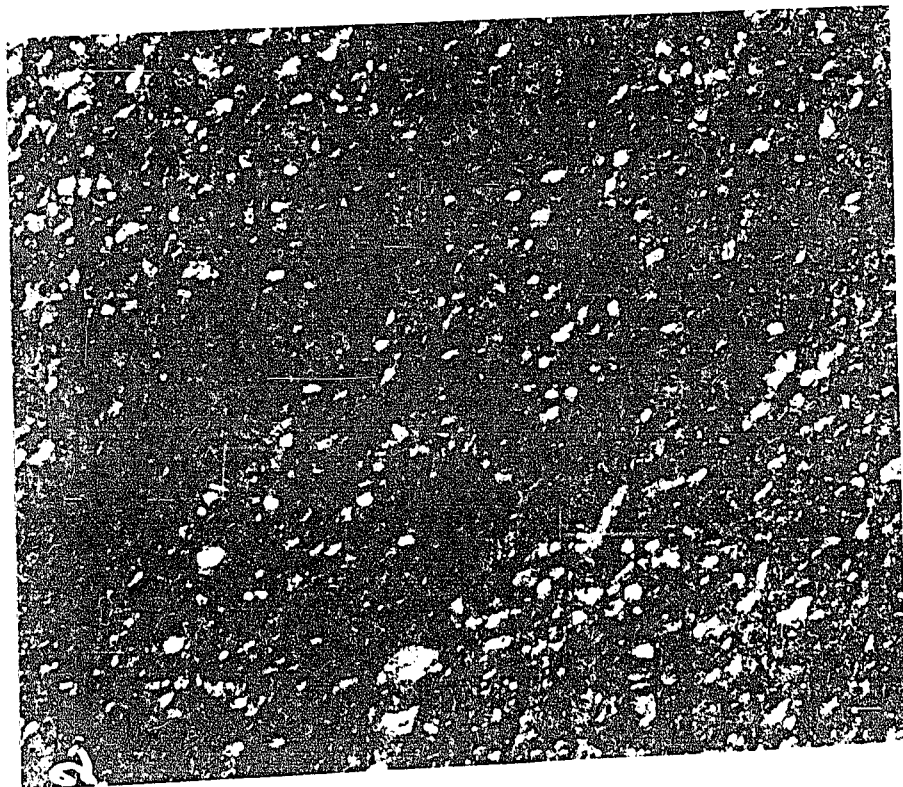
- a) Pseudo-bedding parallel to S_1 cleavage defined by calcite laminae that were formed by segregation (x 50). Phyllitic-slate unit (loc. 432).

- b) Fracture cleavage (S_2) defined by closely spaced parallel fractures. No appreciable mineral alignment along S_2 . The cleavage is at high angle to the bedding (S_0) defined by lighter quartz-rich and darker pelite-rich layers (x 30). Siltstone unit, (loc. 182).

PLATE 9



a)



b)

PLATE 4

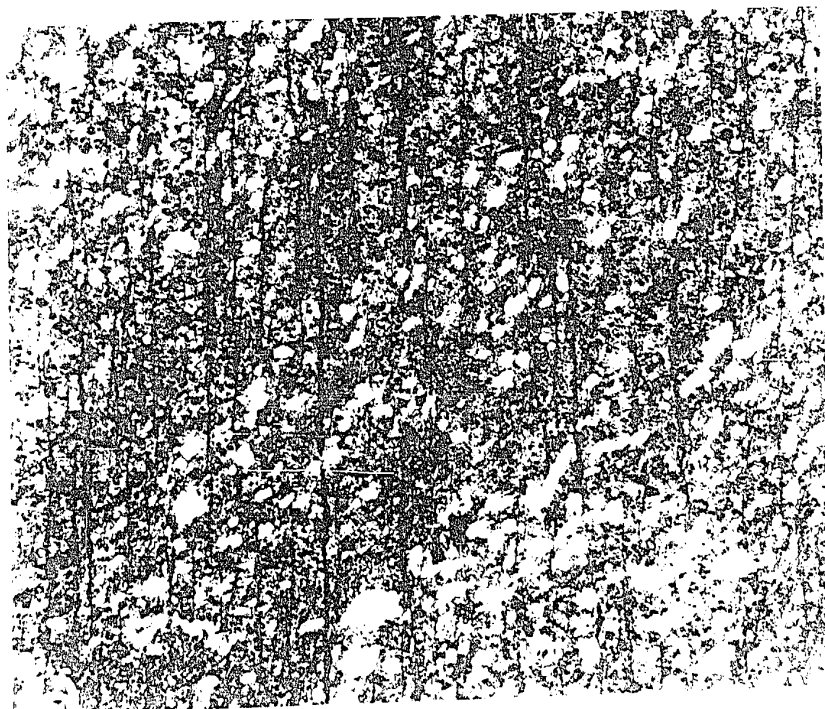
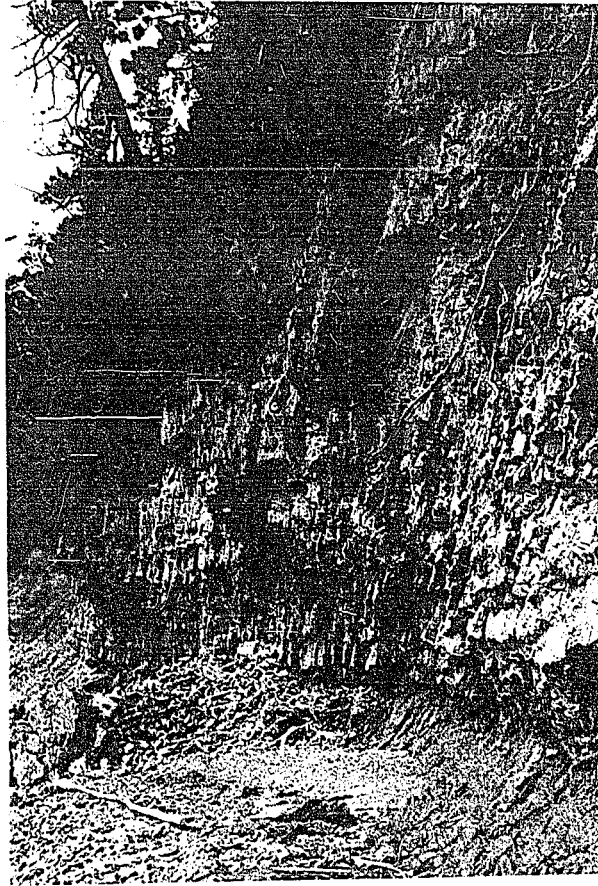


Plate 10 Cleavages in Siluro-Devonian formations of Témiscouata area.

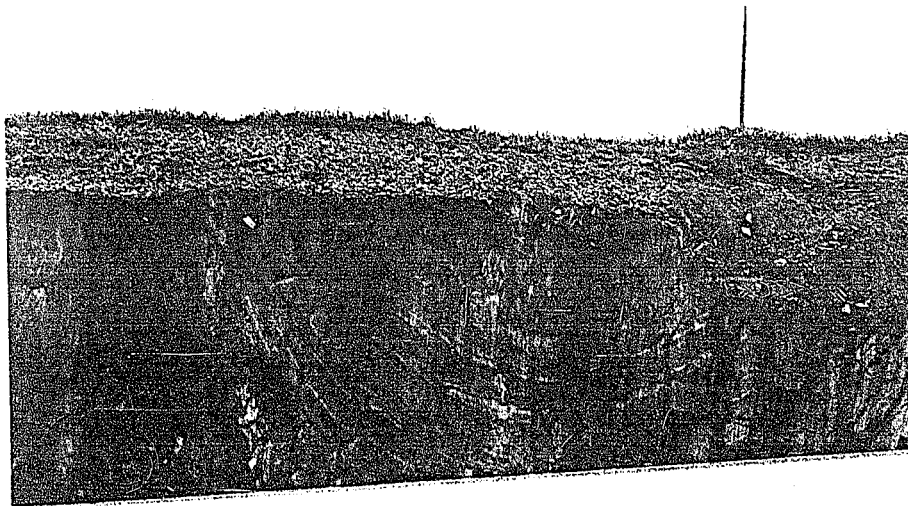
- a) Poorly developed almost vertical fracture cleavage in gently dipping Silurian sandstones (Robitaille Formation) near Témiscouata Lake.

- b) Macroscopically penetrative "fracture" cleavage in gently folded shaly calcareous siltstones of the Lower Devonian Témiscouata Formation (10 km east of Cabano).

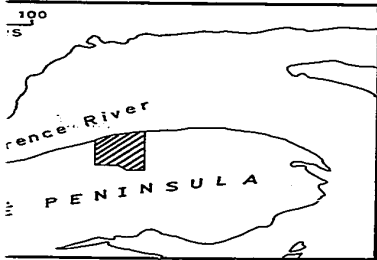
PLATE 10



a)



b)

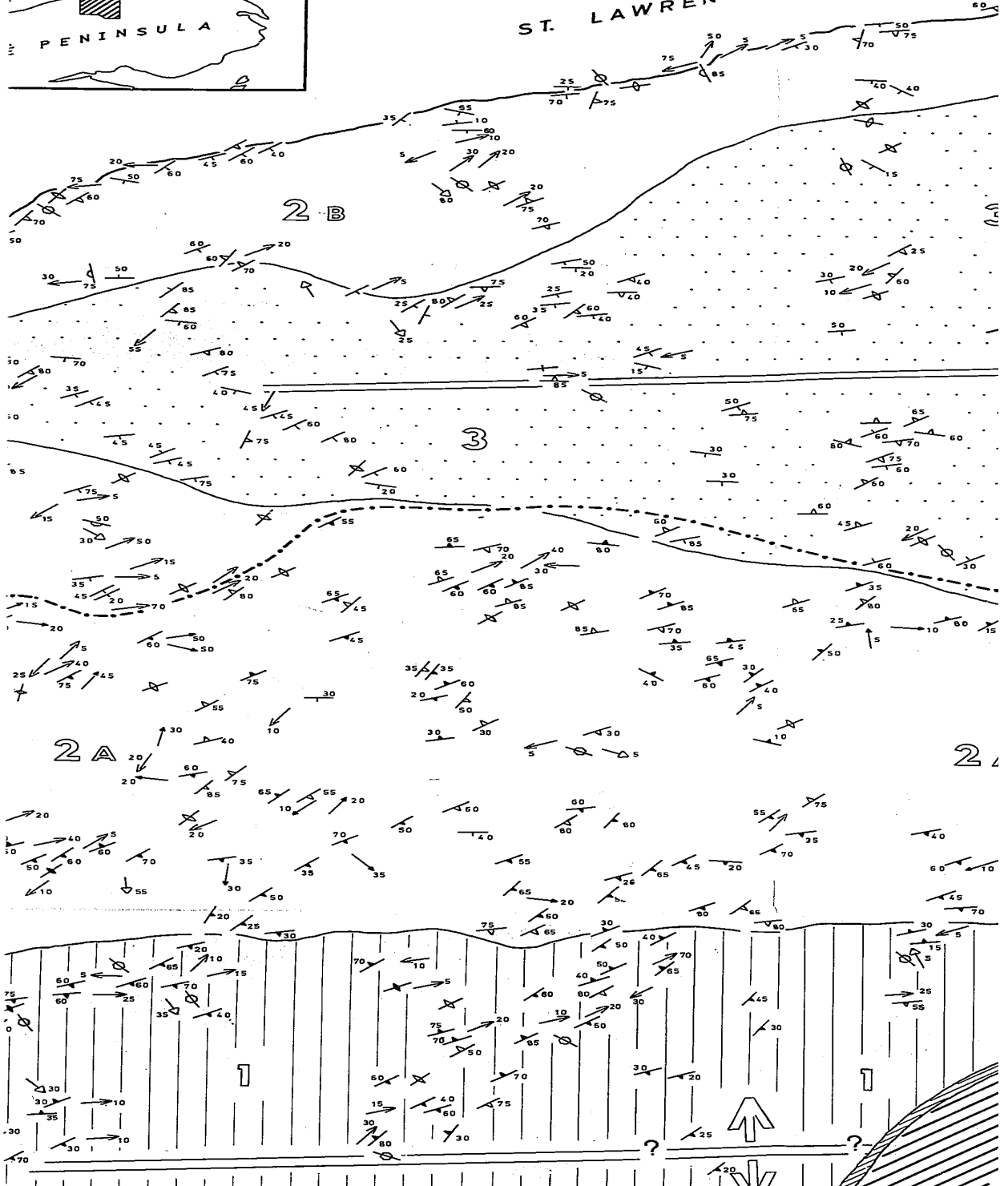


1 of

66°15'

0 L

ST. LAWRENCE RIVER



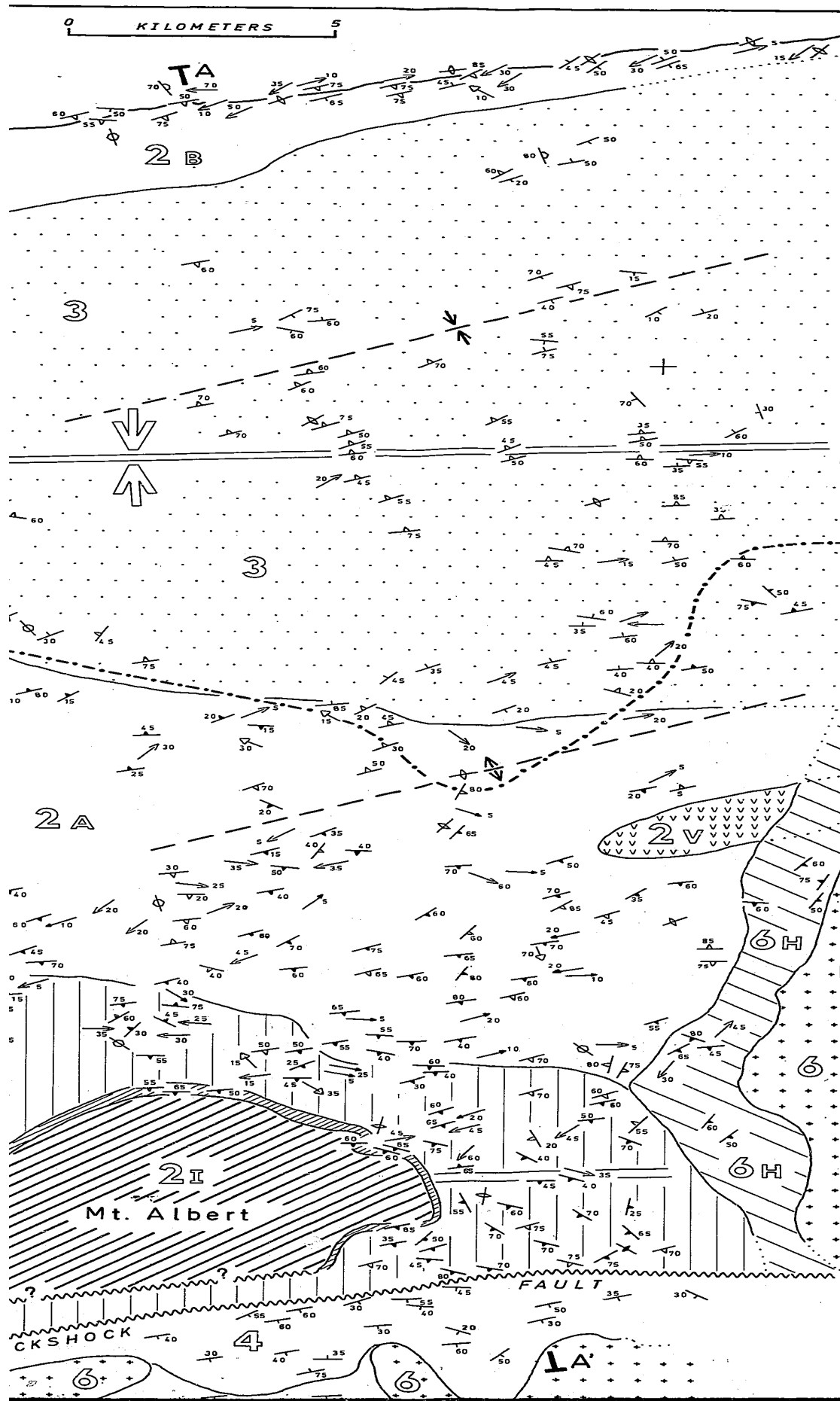
SYNTHESIS

- Granite with Contact Aureole
- Sandstone, Limestone, Shale, Volcanics.
- Sandstone, Limestone, Siltstone, Shale.

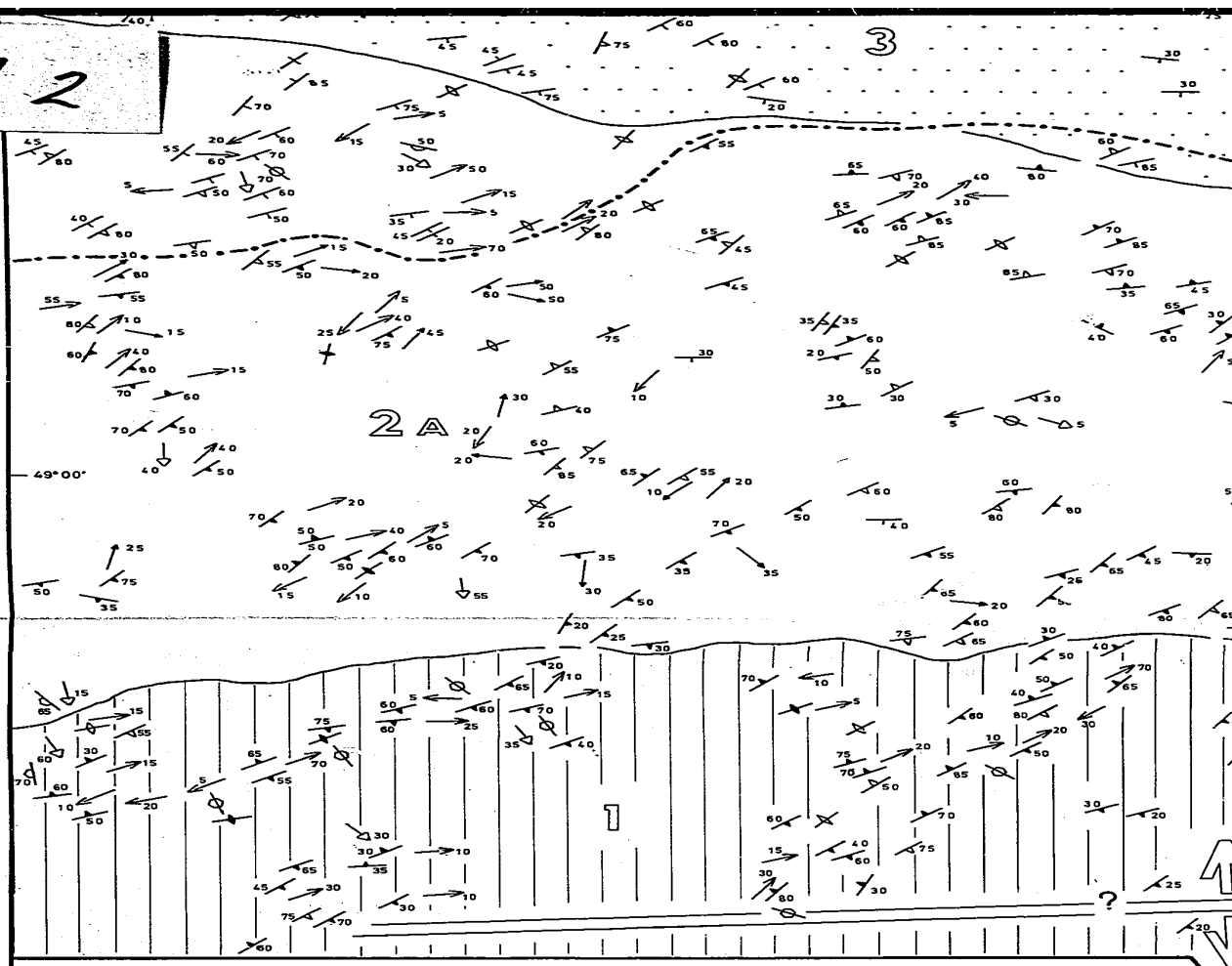
STRUCTURES

- S₀ Bedding
- L₁ Lineation
- S₁ Foliation
- L₂ Lineation
- S₂ Foliation
- L₃ Lineation
- S₃ Foliation

SHICKSH



2 of 2



STRATIGRAPHY

DEVONIAN		Granite with Contact Aureole	
		Sandstone, Limestone, Shale, Volcanics.	
SILURIAN		Sandstone, Limestone, Siltstone, Shale	
MAINLY MIDDLE ORDOVICIAN		Siltstone, Shale Sandstone	Quebec Group
MIDDLE/LOWER ORDOVICIAN		Serpentinite and Contact Aureole	
MAINLY LOWER ORDOVICIAN		Argillite, Greywacke, Limestone, Siltstone	Quebec Group
		Slate, Phyllitic Slate, Limestone, Sandstone, Volcanics.	
CAMBRIAN (?)		Metavolcanics, Minor Metasediments	Shickshock Group

STRUCTURES

	S ₀ Bedding	
	L ₁ Lineation	
	S ₁ Foliation	
	L ₂ Lineation	
	S ₂ Foliation	
	L ₃ Lineation	
	S ₃ Foliation	
		Maj F
		Aca

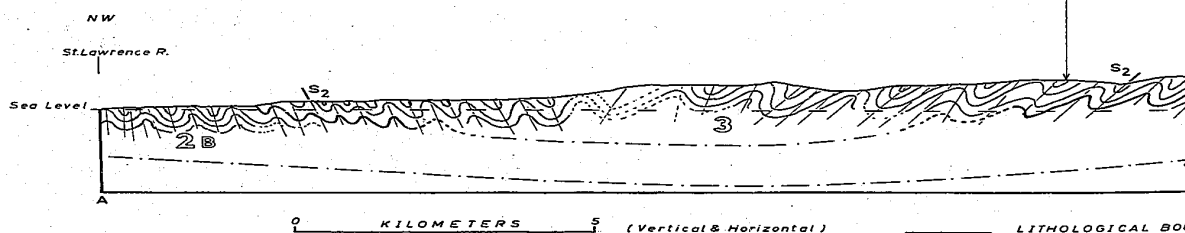
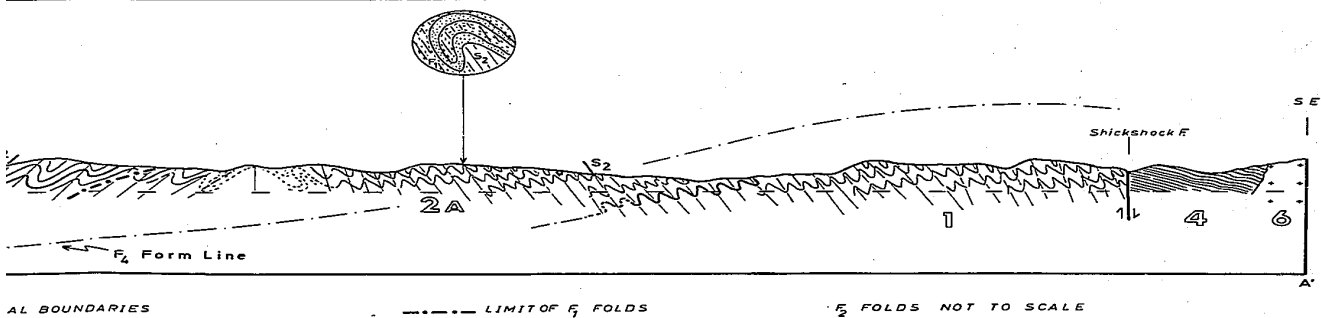
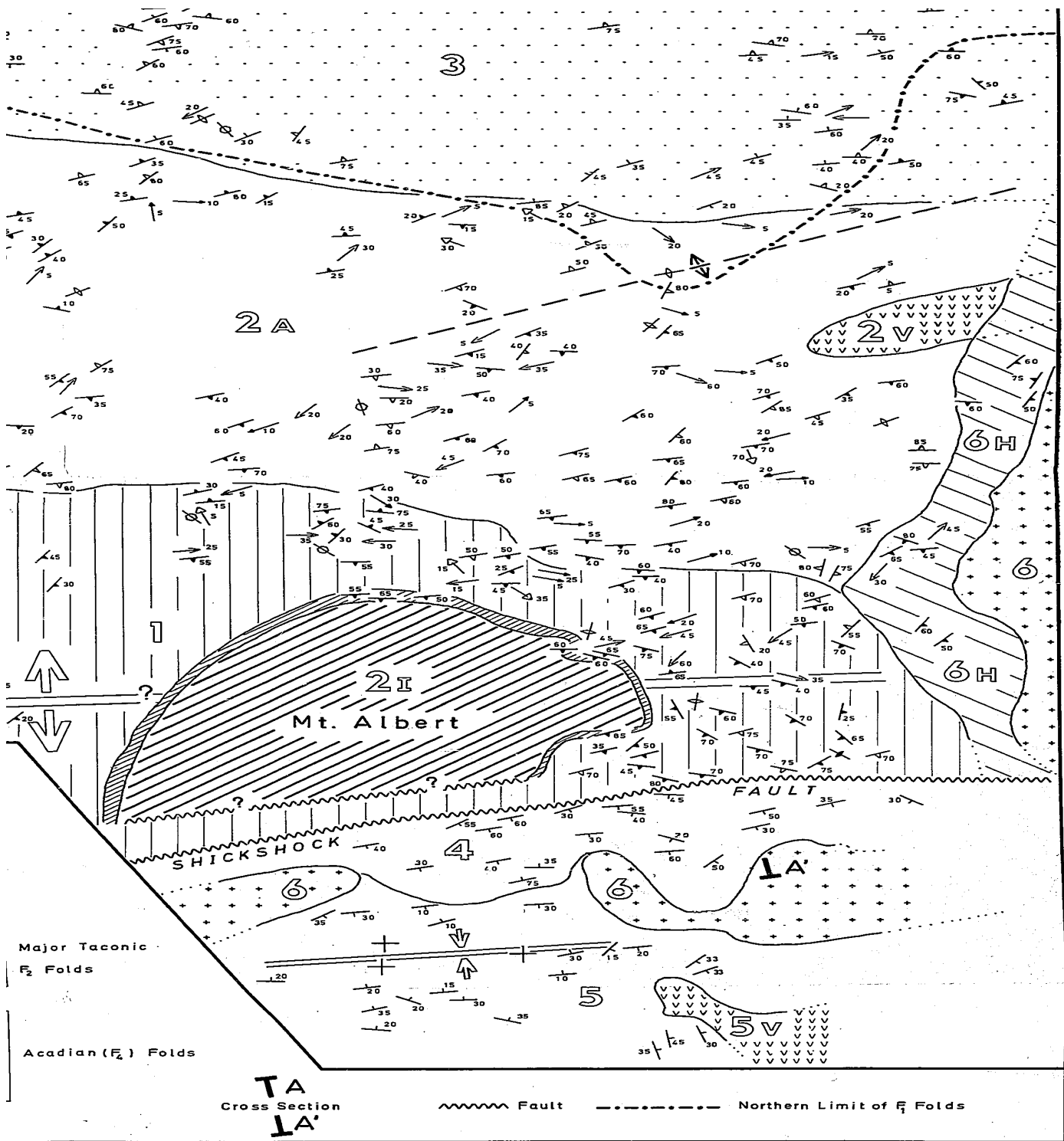


Fig.4. Geological map and interpretative cross-section of Mt. Albert area. Map r



Map modified from McGerrigle (1954), MacGregor (1962), Robert (1966), and Girard (1967).