Geology of the Lewisporte/Loon Bay Area, Newfoundland, Canada

A thesis presented to the Faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of
Master of Science

School of Arts and Sciences
Department of Geological Sciences

Richard F. Livaccari
1980

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Abstract

The Lewisporte/Loon Bay area is located within the Central Volcanic belt of Newfoundland. It contains rocks of (?)upper Cambrian to Ordovician in age. The most important group of rocks within this area is the three part Campbellton sequence. The oldest unit of the Campbellton sequence is the Loon Harbour Formation (500-1,000 m thick) composed of mafic volcaniclastics that is conformably overlain by manganiferous cherts of the Luscombe Formation. The Luscombe Formation (370 m thick) is composed of manganiferous cherts that grade upward into highly argillaceous chert. Conformably overlying the Luscombe Formation is the Riding Island greywacke that represents the uppermost unit of the Campbellton sequence. Other units found within the map area include the New Bay Formation, Dunnage melange, Burnt Bay chert (new), Caradocian age black slate, Goldson Formation and the Botwood Group. The Campbellton sequence is interpreted to underlie the Dunnage melange with the Riding Island greywacke representing a member of the New Bay Formation. The Burnt Bay chert (new) is inferred to directly underlie the Caradocian age black slate and overlie both the New Bay Formation and Dunnage melange. Greywackes correlative with the Sansom/Point Leamington greywackes are not found within the map area.

The dominant structure of the Lewisporte/Loon Bay area consists of a series of moderate to steeply southeast-plunging inclined to reclined, close to tight, overturned macroscopic folds. Axial surface cleavage of these folds forms a regional penetrative cleavage that affects all rocks within the map area. These folds are

interpreted to represent the first major phase of deformation that affected this area (D_1). Various local complications exist such as minor differences in style and orientations of the folds and associated lineations and cleavage. Some of the minor changes in fold orientations may be a result of a series of north-northeast trending sinistral faults that kink and offset cleavage. Important non-penetrative soft sediment structures pre-dating the major regional folding episode are found in several units.

The Carmanville melange may represent the accretionary prism of a west dipping subduction zone of the central volcanic belt of Newfoundland during the late Cambrian to medial Ordovician. Deposition of the Luscombe Formation most likely occurred in the forearc basin of this arc system during its incipient development as nearby arc-related subsea volcanism pumped large quantities of Mn, Fe and Si into the sea water to be precipitated as manganiferous chert. Development of this arc system through time lead to the deposition of the New Bay Formation and Lawrence Head volcanics adjacent to the forearc trough resulting in gravitational slope instabilities and the olistostromic deposition of the Dunnage melange in this forearc trough.

Acknowledgments

I am grateful to W.S.F. Kidd for introducing me to this problem and for his continued support and guidance. This fieldwork was supported by the National Science Foundation Grant #EAR 7708653 awarded to W.S.F. Kidd. I also thank Donna Kelly for typing this thesis.

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CHAPTER I

INTRODUCTION

The Loon Bay/Lewisporte area is located along the north-central coast of Newfoundland, Canada. Topography in the mapped area is subdued with less than 200 feet of vertical relief inland. Outcrop along the coast is quite good but unfortunately lacking in some critical areas. All of the coastal exposure can be reached on foot requiring minimal use of a boat.

July and August of 1978 was spent mapping in the field. Also, a few days in July of 1979 were spent within the field area. Mapping was done on air photos at a scale of 1:17,000 and a geological map of the same scale was constructed (Plate I).

The map area lies within the Central Volcanic Belt or Dunnage zone of Newfoundland that is sandwiched between Precambrian zones of the Long Range Mountains to the east and Avalon Peninsula to the west (see Bird and Dewey, 1970; Williams, 1979). Rocks in this region consist mainly of mafic to rare silicic volcanics and their detritus underlain by an upper Cambrian/lower Ordovician oceanic basement. Wilson (1966) was the first to recognize that the broad regional pattern of stratigraphic/tectonic assemblages in Newfoundland most likely represented the preserved remnants of the opening and closing of an early Paleozoic ocean. Bird and Dewey (1970) and Dewey and Bird (1971) further refined the evolution of Newfoundland within the context of a plate tectonic model. They claimed that central Newfoundland represented the vestiges of an Ordovician island arc complex that could be grossly subdivided into three zones--the Roberts Arm/Lushes

Bight terrain, the Exploits Terrain and the Dunnage melange zone (Fig. 1-1). The Lewisporte/Loon Bay area is located within the Dunnage melange zone. This area lies just south of the southeast extension of the Dunnage melange and is partially separated from the Dunnage melange by the Loon Bay batholith (Fig. 1-2).

Kranck (1952) and Williams (1963) produced general geological maps that included the Lewisporte/Loon Bay area. Kay (1975) first recognized that an important unit of manganiferous chert was located just south of the Dunnage melange in the Loon Harbour/Campbellton area. Kay (1975) named these cherts the Luscombe Formation and described the stratigraphic relationships of this unit with the adjacent rocks. He concluded that the Luscombe Formation was conformably underlain by the Loon Harbour volcanics and conformably overlain by the Riding Island greywacke. Collectively, Kay (1975) named these three units the Campbellton Sequence. What brought the most attention to the Campbellton Sequence was the fact that Kay (1975) claimed that the manganiferous cherts of the Luscombe Formation must be of abyssal origin and could represent sediments overlying an ophiolite. If this is correct, than the Luscombe Formation is of paramount importance to the understanding of the regional geological evolution of central Newfoundland.

The main purpose of this project was to greatly refine Kay's (1975) general map of the Campbellton Sequence and define more precisely the thickness, age and regional extent of the Luscombe Formation as well as to formally define type localities for the units of the Campbellton Sequence which Kay (1975) failed to do. The other major purpose of this project was to define the relationship of the Campbellton Sequence to other nearby units such as the New Bay Formation and

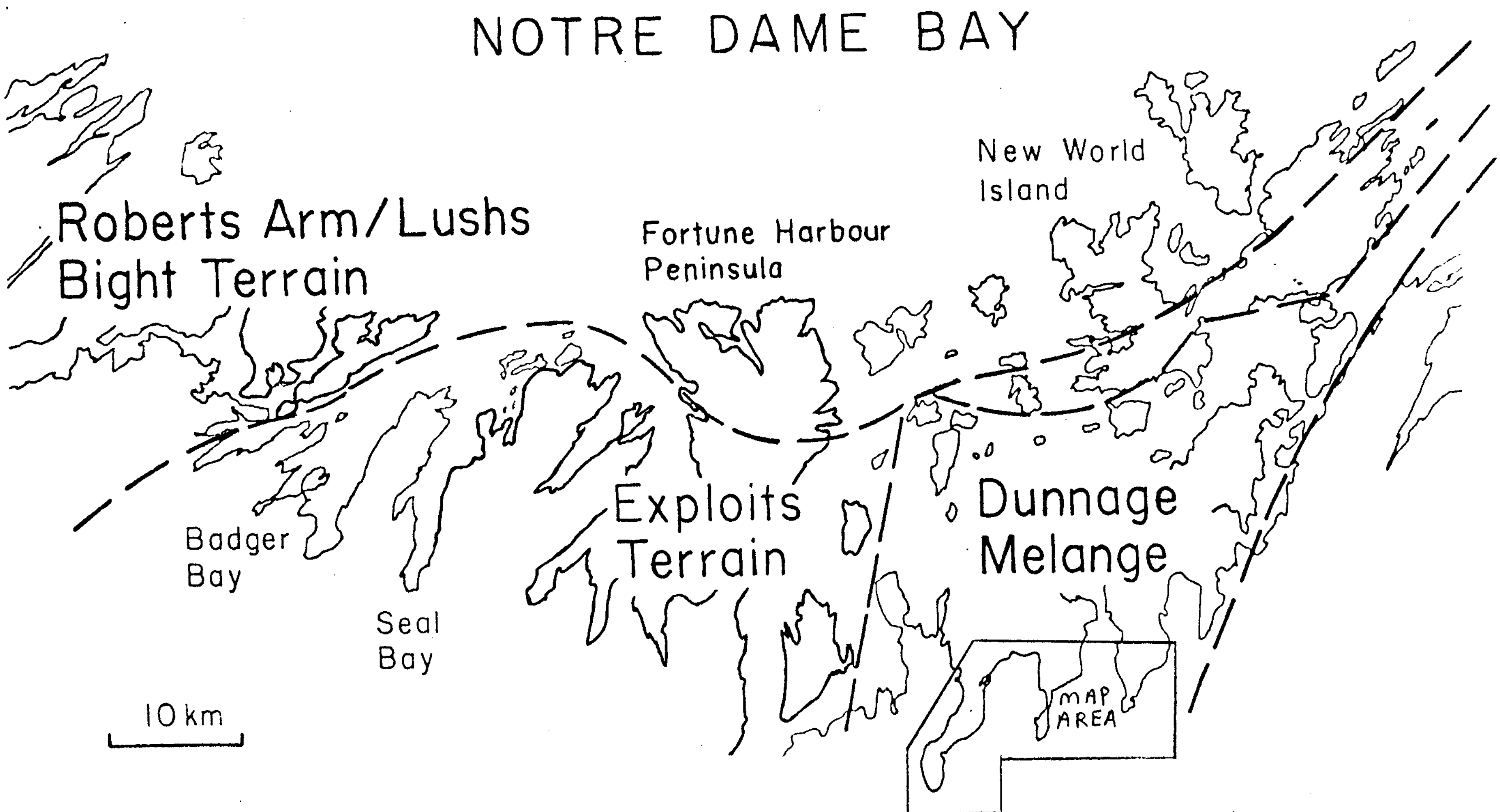


Figure 1-1 (from Nelson, 1979).

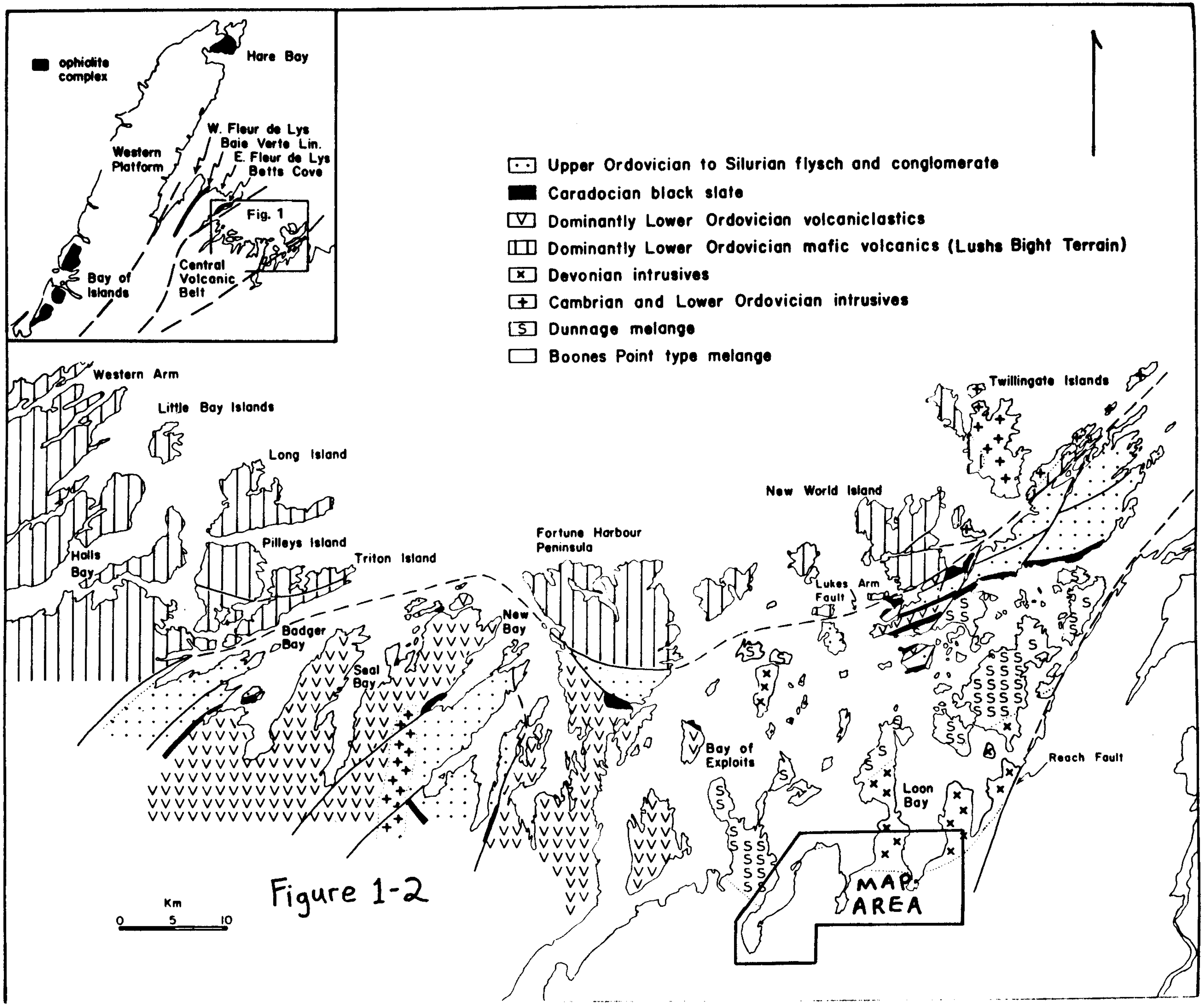

ophiolite complex
Hare Bay
W. Fleur de Lys
Baie Verte Lin.
E. Fleur de Lys
Betts Cove
Western Platform
Fig. 1
Bay of Islands
Central Volcanic Belt
Upper Ordovician to Silurian flysch and conglomerate
Caradocian black slate
Dominantly Lower Ordovician volcaniclastics
Dominantly Lower Ordovician mafic volcanics (Lushs Bight Terrain)
Devonian intrusives
Cambrian and Lower Ordovician intrusives
Dunnage melange
Boones Point type melange
Western Arm
Little Bay Islands
Long Island
Pilleys Island
Triton Island
Halls Bay
Badger Bay
Seal Bay
New Bay
Fortune Harbour Peninsula
New World Island
Lukes Arm Fault
Twillingate Islands
Bay of Exploits
Loon Bay
Reach Fault
MAP AREA
Km
0 5 10

Figure 1-2

Dunnage melange.

CHAPTER II

LITHOLOGY

Sedimentary Rocks

Loon Harbour Formation

This unit was originally defined by Kay (1975) for outcrops of volcaniclastic rocks exposed along the shores of Loon Harbour and in the woodland area south of Campbellton. Kay (1975) named this unit the Loon Harbour volcanics but did not specify a type locality for this unit. Since this unit is well exposed, I propose that it should be raised to formational status whose type locality are exposures of volcaniclastic rocks along the southern half of Luscombe Point in Loon Harbour and along the western shore of Loon Harbour south of the Luscombe Formation.

A Mesozoic age stock of lamprophyre intrudes the Loon Harbour Formation south of Dildo Pond (see Williams, 1964). Kay (1975) misinterpreted this lamprophyre body as being part of the Loon Harbour Formation. This mistake lead Kay (1975) to believe that the Loon Harbour Formation represented the upper portions of an ophiolite. This is quite incorrect since this unit consists almost entirely of massive, drab-green, fine-grained to pebbly fragmental volcaniclastic rocks.

Outcrops of the Loon Harbour volcanics are confined to an east-west trending zone between Loon Harbour and Campbellton. Good exposures of the Loon Harbour volcanics are found south of the Luscombe Formation around the shores of Loon Harbour, in road cuts along route 340 and southward in the woodland area. Good outcrops are also found south of the section near the old mill in the woods south of Campbellton.

Poorly defined bedding in the Loon Harbour volcanics masks any large scale internal structures making the thickness of this unit difficult to estimate. Outcrops of the Loon Harbour volcanics south of the overlying subvertical Luscombe Formation in Loon Harbour stretch 2.5 km southward perpendicular to strike and stretch 0.75 km south of the Luscombe Formation south of Campbellton. The bottom of this unit is not seen since it is cut off to the south by a fault dropping Goldson conglomerate against the Loon Harbour volcanics. The section south of Campbellton appears to be bereft of any large scale folding and therefore the exposed thickness of this unit is a minimum 500 m as measured on the map in subvertical strata south of Campbellton. However, this unit may be several km thick as measured on the map along sparse woodland outcrops south of Loon Harbour, proper.

The lower portion of the Loon Harbour volcanics is in fault contact with the Silurian Goldson conglomerate. The upper contact of the Loon Harbour volcanics is a rapidly, gradational conformable contact with the Luscombe Formation (discussed in more detail below).

Excellent exposures of the upper part of this unit are seen in Loon Harbour. Along the west side of Loon Harbour the contact with the overlying Luscombe Formation is exposed. Near this contact the Loon Harbour volcanics consists of interbedded volcaniclastics, minor chert and minor coarse-grained wackes. The volcaniclastics are fine to medium grained and dirty green in color. Beds are graded and vary from 4 cm to 2 meters thick. Thin sections show these volcaniclastics to consist of 20% highly altered medium-grained plagioclase, 5% medium-grained mafic volcanic clasts and 75% mostly recrystallized matrix of dominantly muscovite, carbonate and chlorite with lesser amounts of sphene and epidote.

Interbedded wacke layers are graded in 5-30 cm thick beds. They consist of very coarse to fine-grained sand size particles and are pinkish-white in color. Thin sections show these wackes to consist of up to 60% coarse to medium-grained, angular plagioclase that sometimes exhibits pericline twinning and is somewhat sericitized, up to 15% sub-angular, coarse-grained, quartz clasts set in a recrystallized matrix of quartz, epidote, calcite, sphene and opaque pyrite.

The intercalated chert layers found between the volcaniclastic and arkosic wacke beds in this Loon Harbour volcanics/Luscombe Formation transition zone consist of green argillaceous chert in beds up to 20 cm thick. Thin sections show this chert to be composed of 50% microcrystalline quartz, 30% very fine-grained chlorite, 5% spherulitic quartz forms and 5% euhedral to anhedral fine grains of spessartite.

Good exposures of the uppermost part of the Loon Harbour volcanics are seen on the east side of Loon Harbour on the southern half of Luscombe Point. Here pillow basalt blocks several meters across are found in a matrix of dark gray, fine-grained to pebbly mafic volcaniclastics. The pillows within the blocks are cylindrical in shape up to 2 by 2 meters wide and several meters long and consist of gray vesicular mafic lava with radially elongated pipe vesicles up to 2 cm across. Vesicles in some cases make up about 50% of the total rock. A very fine-grained greenish gray chlorite rind about 1 cm wide mantles each pillow. Thin, wispy, yellow-gray limey to argillaceous material up to 2 X 30 cm across fills triangular voids between pillows. These pillow basalt blocks are sparse and constitute less than 10% of the total outcrop.

The dominant rock type near the Loon Harbour/Luscombe Formation contact are gray, fine-grained to pebbly volcaniclastics. The basaltic pebbles are up to 10 X 20 cm across, averaging 1 X 5 cm across and are consistently elongated (tectonically) down-dip. They are composed of gray vesicular to amygdaloidal pillowed mafic lava fragments set in a fine-grained calcite-rich (mafic) matrix. Amygdules consist of white crystalline calcite and compose up to 50% of the rock. Thin sections show the mafic lavas to consist of 40% very fine-grained groundmass, 30% chlorite, 20% epidote and 10% sphene.

Pillow basalt blocks are also found in the Loon Harbour volcanics in road cuts just west of Loon Harbour along route 340. Here elongate pillows up to 1 1/2 m long are found in a 20 m wide zone. The pillows are gray-green weathering to a dull green and are amygdaloidal to vesicular. The pillow basalt blocks are surrounded by plagioclase and calcite-rich, medium to fine-grained, gray-green mafic (probably basaltic) volcaniclastics. Rare rounded rip-up clasts 1-3 cm across are found within these volcaniclastics. More weathered zones are drab powdery green in color with sporadic red-brown stains.

Woodland outcrops of the Loon Harbour volcanics south of Loon Harbour are comprised of monotonous, massive to medium bedded (10 cm to 1 m beds), drab green mafic volcaniclastics. Grain size varies from fine-grained sands to pebbles. Conglomerate zones consist of angular to sub-rounded pebbles up to 20 cm across composed of vesicular to scoriaceous mafic lava, and pillow fragments set in a drab green sandy fine- to coarse-grained, poorly sorted volcaniclastic matrix. This matrix makes up between 30-50% of the conglomerate. Up to 10% of the matrix consists of pink to white crystalline calcite. Conglomerate beds are massive to medium-bedded in 1 to 1/2 m thick beds. Bedded zones are graded with scour marked and load casted bases. Other outcrops consist of massive to medium-bedded, silty to sandy, vesicular to cherty looking gray to green volcaniclastics. These cherty volcaniclastics are sometimes graded and, in places, are made up of coarse to fine sand size clastic plagioclase and altered pyroxene grains set in a cherty matrix. Soft sediment slump structures are rarely seen in the fine-grained cherty volcaniclastics. Thin sections show some of the fine-grained zones to consist dominantly of secondary sphene, chlorite, epidote and altered plagioclase with lesser amounts of talc and calcite.

South of Campbellton, the Loon Harbour volcanics consist mainly of sandy very fine to medium-grained, dull green to dark gray, massive, cherty volcaniclastics. Thin sections of the cherty samples are composed of 80% extremely fine-grained chert/chlorite matrix surrounding angular to subrounded fine clastic grains of quartz, sericitized plagioclase and porphyritic fine-grained mafic volcanic fragments. Thin sections of the non-cherty volcaniclastics consist of up to 50% very fine-grained aggregates of recrystallized chlorite, sphene, epidote and chert, 40-30% highly altered subrounded to broken angular plagioclase laths, up to 15% angular quartz clasts, and a few percent of hornblende, augite and porphyritic mafic volcanic clasts.

Luscombe Formation

The Luscombe Formation was originally defined by Kay (1975) as the middle member of the tripartite Campbellton Sequence for outcrops between Loon Harbour and Campbellton. As in the case of the Loon

Harbour volcanics Kay (1975) did not formally propose a type locality for this formation. Therefore I propose that the type section are outcrops above the Loon Harbour volcanics located along the northern portion of Luscombe Point in Loon Harbour.

In general, the Luscombe Formation consists of gray to green argillaceous cherts whose argillite component increases both up section and westward from Loon Harbour towards Campbellton (Figs. 2-1, 2-2). A very important manganiferous chert zone is found in the center of the Luscombe Formation that consists of layers rich in spessartite and rhodochrosite interbedded in the chert.

The Luscombe Formation is best exposed around the shores of southernmost Loon Bay and along a dirt path south of the old mill in Campbellton. Sections were measured in both places. Good but widely spaced road cuts of the Luscombe Formation are found along route 340 between Campbellton and Loon Harbour. Woodland outcrops east and west of Loon Harbour are quite good but become very sparse further inland. Good outcrops of the Luscombe Formation are also found along the shore north of Campbellton, just south of Rideout Point.

The lower contact of the Luscombe Formation with the underlying Loon Harbour volcanics is exposed along the southwestern shore of Loon Harbour where it is seen to be conformable and gradational over a thickness of about 10 meters. This zone consists of interbedded fine-grained to pebbly, graded volcaniclastic to quartz/plagioclase bearing wacke beds 1.5 to 5 m thick interbedded with gray-green chert. The base of the Luscombe Formation was taken to be at the top of a 5 m thick pebbly to very fine-grained volcaniclastic bed above which chert dominates.

Generalized Stratigraphic Column of Luscombe Formation in Loon Harbour Area

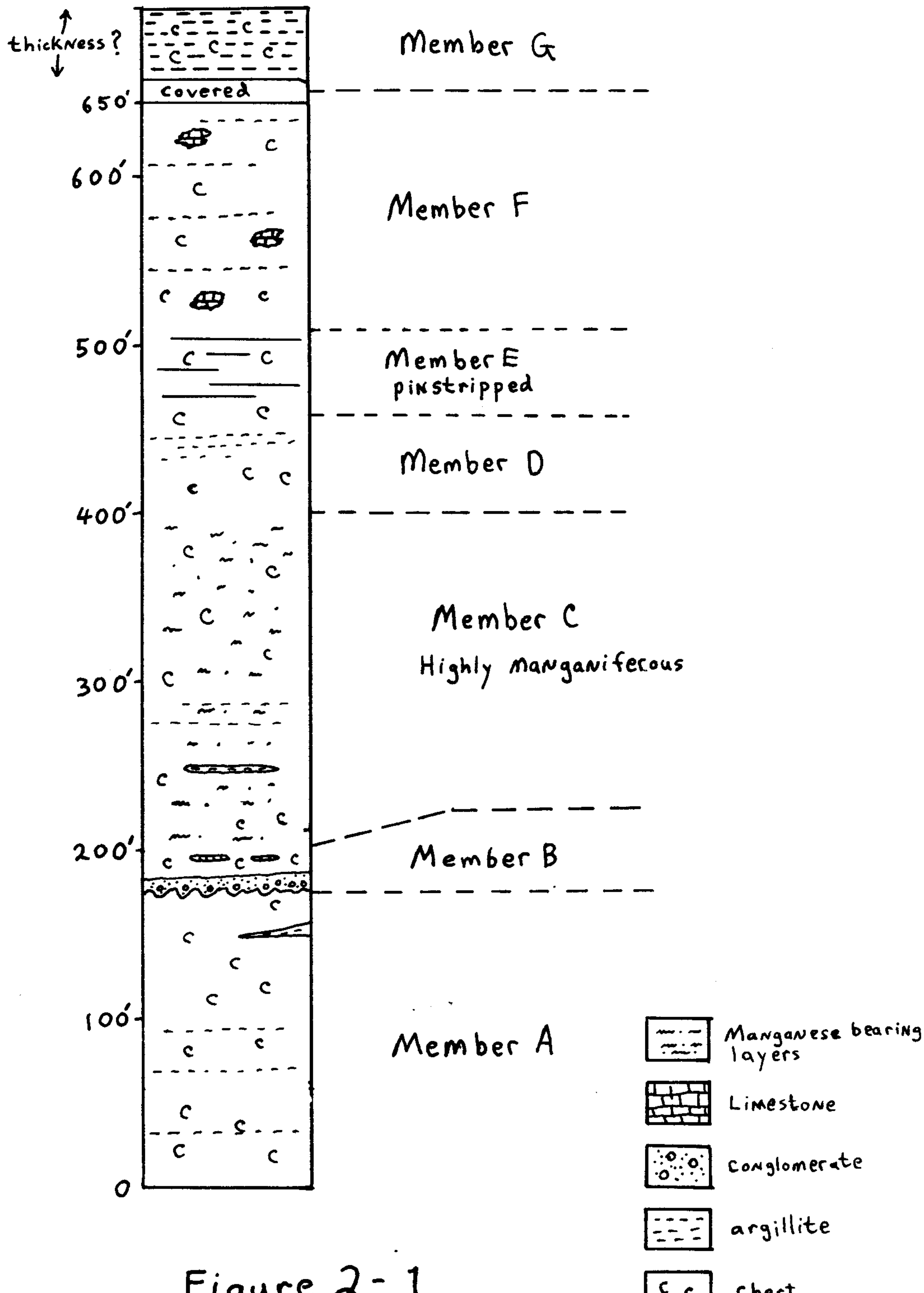


Figure 2-1

Generalized Stratigraphic Column of the Luscombe Foration in the Campbellton Area (old Mill section).

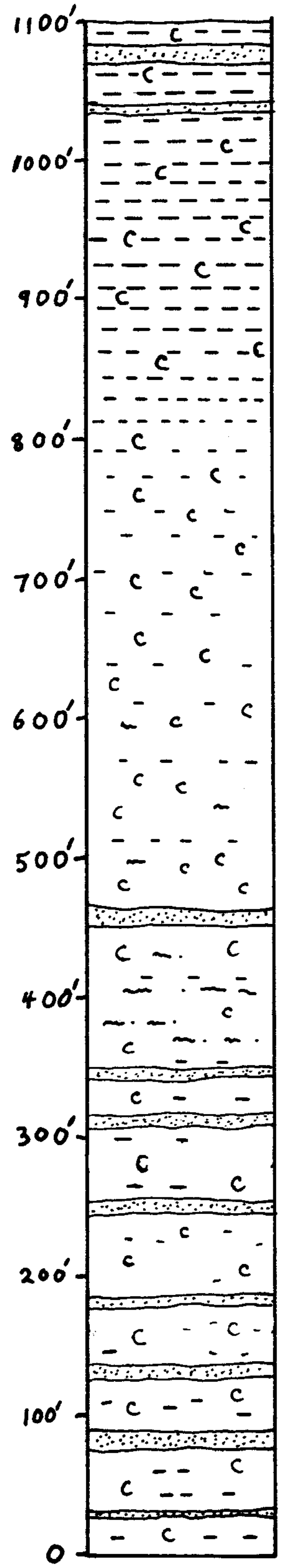


Figure 2-2
(Same key as figure 2-1)

The measured thickness of the Luscombe Formation is about 370 m (1100 ft.; see plates III and IV). Detailed stratigraphic sections were measured in the Loon Harbour and Campbellton areas. These sections are presented in plates III (Loon Harbour section) and IV (Campbellton section) and described below. Figures 2-1 and 2-2 summarize these detailed straigraphic sections.

The Luscombe Formation has been divided into 7 members (A through G) on the basis of variations in lithology. These 7 members are well defined along the shores of Loon Harbour but lose their distinctive characteristics further west towards Campbellton (Fig. 2-1). A detailed description, by member, of the Luscombe Formation at southern Loon Bay goes as follows (see also Plate III):

Member A - Member A is the lowermost member of the Luscombe Formation and is approximately 50 m (170 ft) thick. It consists of purple, gray and green chert (Fig. 2-3). Very fine-grained mica spangles (metamorphic, not clastic) can be discerned in hand specimens. Thin sections reveal these mica spangles to be chlorite making up 20% of the rock with crypto- to microcrystalline recrystallized quartz and a few percent of opaque fine-grained pyrite and haematite composing the remainder of the rock. Traces of fine-grained detrital quartz, plagioclase and epidote were also found in thin sections. The chert is bioturbated with its argillite (chlorite) component increasing upward. Interbedded green, argillaceous, mafic volcaniclastic layers (?tuff) 1-20 cm thick are common. The lower half of member A contains rare thin (up to 5 cm thick) graded quartz-plagioclase-rich, coarse-grained wacke layers.

In the lower part of the exposed portion of member A on the western

Figure 2-3. Member A of the Luscombe Formation

shore of Luscombe Point a soft-sediment, low-angle thrust fault is seen. This fault cuts through less than a 40 cm thickness of beds offsetting them by about 10 cm. Beds above and below this thrust fault are undisturbed.

Member B - Member B is 9 m (30 ft) thick and is composed of 3 units. The lowest unit is a 3.2 m (10') thick graded, fine-grained to pebbly, quartz-plagioclase rich volcaniclastic bed with a scour marked and load casted base. The lower part of this bed contains pebbles and chert rip-up clasts up to 15 cm across. Above the volcaniclastic bed is a gray to green chert unit with thin (1-3 cm) argillaceous interbeds similar to Member A. This passes upward into the uppermost unit of Member B that consists of dark gray argillaceous chert with interbedded buff limestone lenses up to 4-5 cm thick and 1 m long (Fig. 2-4). Manganese nodules and lenses of rhodochrosite and spessartite up to 2 by 10 cm are also found in this uppermost part of Member B. Thin sections show the buff limestone lenses to consist of nearly pure crystalline calcite mantled by a thin spessartite rich chert zone. These limestone lenses have a pinch-and-swell appearance suggesting they once formed continuous beds that have been diagenetically dissolved away.

Member C - Member C is the distinctive manganiferous rich member of the Luscombe Formation and is 56 m (185 ft) thick. The lower half of this member consists of bioturbated dark maroon to dark gray-green chert containing up to 50% interbedded manganese nodules and lenses of rhodochrosite and spessartite that vary in size from 1-20 cm long and 2-4 cm thick. Interbedded in this chert are rare green intermediate or mafic volcaniclastic (?tuff) beds up to 10 cm thick (Fig. 2-5).

Figure 2-4. Limestone lenses in Member B of the Luscombe Formation.

Figure 2-5. Probable volcanic tuff layers in Member C of the Luscombe Formation.

The upper half of Member C differs from the lower half in that it is comprised of alternating zones of manganiferous and non-manganiferous chert (Fig. 2-6). The non-manganiferous chert zones consist of homogeneous dark gray-green chert whereas the manganiferous zones are composed of up to 50% nodules and lenses of rhodochrosite and spessartine in gray-green chert. These alternating layers vary from 20 cm to 1 m thick.

Thin sections show that the homogeneous chert consists of up to 20% very fine-grained chlorite, 70% microcrystalline quartz and 10% oval to round granular quartz aggregates less than 1 mm across.

The manganiferous zones are complex in thin section. They consist of a mixture of dominantly crystalline rhodochrosite and very fine-grained, euhedral spessartite with lesser amounts of quartz, epidote, biotite and secondary actinolite. In these zones the spessartite and rhodochrosite are either segregated or homogeneously mixed. Other pink manganiferous layers consist chiefly of very fine-grained euhedral spessartite crystals with minor amounts of quartz, biotite and actinolite. Gray chert layers that separate the pink manganiferous layers contain up to 20% radiating clusters of actinolite and very fine-grained euhedral spessartite crystals. Other gray chert layers separating the pink manganiferous zones are composed of 70% micro-crystalline chert and 30% very fine-grained chlorite with only a trace of very fine-grained spessartite. Thus, there appear to be two types of manganiferous layers as follows:

1) Nodules and lenses comprised mainly of subequal amounts of spessartite and rhodochrosite.

Figure 2-6. Upper portion of Member C of the Luscombe Formation.

Figure 2-7. Manganese rich layers in Member C of the Luscombe Formation.

These nodules and lenses are pink on a fresh surface and weather to a jet black mixture of pyrolusite and manganite. They easily weather out of the surrounding more indurated gray chert leaving a "black hole" appearance.

2) Thin, wispy more laterally continuous layers usually less than 1 cm thick of pink-gray material that mantle the black weathering spessartite and rhodochrosite lenses and nodules (see Fig. 2-7). These layers contain primarily very fine-grained euhedral spessartite crystals with subordinate amounts of quartz, chlorite and secondary actinolite. These lenticular and wispy, pink-gray layers weather to a whitish-pink but do not easily weather out of the surrounding chert.

Manganiferous zones exhibit pinch-and-swell structure with the thicker, black-weathering rhodocrosite and spessartite lenses and nodules pinching out and giving way to the thinner lenticular spessartine bands that mantle the lenses and nodules. This is very likely the result of differential diagenetic dissolution of once more continuous manganese carbonate layers leaving behind sufficient amounts of manganese to allow the later metamorphic growth of spessartite in the thin, wispy pink layers.

The presence of gray chert layers between the pink manganiferous layers suggests cyclic deposition of manganese-rich carbonate and chert. Non-penetrative soft sediment folds (F_o) are very well developed and commonly seen throughout the thickness of Member C (Fig. 2-8). These folds are gentle to tight with wavelengths of 10 to 30 cm. They are usually bounded above and below by undisturbed beds. In the upper portion of Member C along the western shore of Loon Harbour a layer approximately 1 meter thick exhibits a chaotic pattern of tight to

Figure 2-8. Open soft sediment folds in Member C of the Luscombe Formation.

isoclinal polyclinal folds of soft sediment slump origin.

Member D - Member D is 14 m thick (45 ft.) and consists of homogeneous dark gray bioturbated chert. Thin sections show this chert to be composed of up to 30% very fine-grained chlorite and/or biotite, 70% microcrystalline chert and a few percent of round polycrystalline, quartz aggregates. Rarely interbedded with the chert are friable green argillaceous mafic to intermediate volcaniclastic layers (?tuff) less than 3 cm thick.

Thus it appears that this member represents a hiatus in manganese-rich sediment deposition.

Member E - Member E is 15 m thick (50 ft.) and is composed of "pinstriped" gray-green chert (Fig. 2-9). The pinstripes are light gray in color, weathering buff and are usually less than 1/2 cm thick. They tend to pinch out over a distance of several meters to less than 1 cm and are spaced 1/2 cm to 1 mm apart. Thin sections show that these pinstripes consist of a mixture of finely crystalline rhodochrosite and very fine-grained euhedral crystals of spessartite. The chert between the pinstripes consists of 70% microcrystalline quartz and 30% chlorite. Chlorite layers tend to cut obliquely across the pinstripes exhibiting a microstylolitic texture. This suggests diagenetic dissolution of once more extensive manganese carbonate layers.

Member F - Member F is more than 51 m thick (170 ft.). It consists of dark gray to green chert interbedded with occasional but distinctive, irregularly shaped micritic limestone pods that are 2-6 cm thick and up to 1 m long (Fig. 2-10). These pods are brown on fresh surfaces weathering out of the chert with a gray to black color. Also interbedded in the chert are thin (2-4 cm thick), green, friable intermediate

Figure 2-9. Member E of the Luscombe Formation.

Figure 2-10. Member F of the Luscombe Formation.

to mafic volcaniclastic layers and rare thin buff-colored spessartite pinstripes.

Thin sections show the limestone pods to consist of up to 20% very fine-grained, euhedral spessartite crystals and about 10% quartz with the remaining 70% consisting of coarsely crystalline carbonate with a slight pinkish tint suggesting that it is composed of manganiferous limestone. Thin sections further show the chert to contain up to 30-40% chlorite, slightly more argillaceous than the chert in the lower members.

Member G - Member G is the uppermost member of the Luscombe Formation. Due to structural complications and incomplete outcrops of this member the thickness is uncertain but postulated to be at least 90 m (300 ft.) thick.

Member G consists of friable pyrite rich, black argillaceous chert to cherty argillite. Thin sections show the rock is composed of subequal amounts of chert and black argillite.

Kay (1975) presented chemical analyses of rocks collected from both Members C and E of the Luscombe Formation. Manganese content of the member C rocks varied from 8-15% with iron contents of 4-7%. Member E rocks contained 4% manganese and 5% iron. Analysis of Member C rocks was also done by S. Brown (Dunn Geoscience Corporation, unpublished data). 'Pink' layers of Member C were found to be composed of 10.6% manganese and 4.2% iron whereas the darker more cherty layers were found to be composed of 4.9% manganese and 4.9% iron. Brown concluded from this data that these rocks do not appear to be of economic interest.

The lithology of the Luscombe Formation in the southern Loon Bay area may be summed up as consisting of up to 300 m (1000 ft.) of gray, green, to black argillaceous chert whose argillite component increases upward from 20% near the base of the section to 50% or more in the uppermost part of the section (Fig. 2-1). The bulk of the manganiferous chert is contained in Member C in the form of interbedded rhodochrosite and spessartite bearing nodules and lenses. Lesser amounts of manganese bearing chert is found in Members E and F in the form of thin, buff spessartite-bearing pinstripes and irregular manganese carbonate lenses, respectively.

A 300 m (1000') section of the Luscombe Formation was measured south of the old mill in Campbellton. Since this section consists of a "dirty" but continuous woodland outcrop rather than along a well washed shoreline outcrop some details of the lithologies are probably obscured. Most of the members distinguished along the Loon Harbour section could not be recognized in the Campbellton section (Fig. 2-2).

The lower half of the Campbellton section (Plate IV) consists of laminated to medium bedded, gray to black, bioturbated argillaceous chert. Bedding is defined by variations in the argillaceous component of the chert. The chert weathers light gray to green with a metallic blue manganese oxide luster to a rusty brown iron oxide coating.

Thin sections reveal distinct manganiferous zones within these cherts that contain a mixture of primarily very fine-grained euhedral spessartite crystals and crystalline rhodochrosite with lesser amounts of chert. Manganese free zones in the chert consist of 80% micro-crystalline (chert) quartz and 20% very fine-grained chlorite with a few percent of fine-grained, spherical, polycrystalline quartz aggregates.

Numerous graded, volcaniclastic layers from thin laminations to beds up to 5 feet thick are found interbedded in the chert throughout the length of the Campbellton section. These beds are gray to brown and very fine to medium grained. Thin sections show these volcaniclastics to contain up to 20% angular quartz grains, 10% sericitized plagioclase, 5% silicic and mafic porphyry clasts with the remaining 65% consisting of very fine-grained cherty groundmass that has predominantly been altered to chlorite.

The upper half of the Campbellton section is composed of gray to black, thin bedded argillaceous chert. This part of the section is rhythmically bedded with indurated argillaceous chert layers alternating with friable cherty argillite layers. These layers vary from 3 to 15 cm thick, are bioturbated and exhibit soft sediment pinch and swell structures. Uncommon thin zones of jet black weathering manganiferous rhodochrosite and spessartite bearing layers 1-2 cm thick are found.

Overlying the thin-bedded chert is friable highly argillaceous pyrite-bearing black chert that weathers rusty brown to sulfur yellow. This black chert is very similar to Member G of the Loon Bay section. Thin sections show this chert to consist of 60% very fine-grained black argillaceous material, 38% microcrystalline quartz and 2% very fine-grained secondary euhedral pyrite. The argillite component of the Campbellton section increases upward from 20% at the base to 60% near the top, like the Loon Bay section.

Outcrop between the Loon Bay and Campbellton sections is sparse and restricted mainly to roadcuts. Roadcuts near the radio tower along route 340 expose manganiferous cherts similar in appearance to Member C of the Loon Bay section exhibiting characteristic pink rhodochrosite and spessartite bands set in dark gray chert. Outcrops around the radio towers expose purple to light green weathering gray color-banded cherts similar in appearance to Member A of the Loon Bay section. Well developed soft sediment convolute fold structures are seen in some of these outcrops. West of the radio tower, towards Campbellton, roadcuts expose manganiferous Member C and pinstriped Member E cherts. Some of the chert in these roadcuts is dark gray in color with a distinctive pinkish tint suggestive of finely disseminated spessartite. In the woodland area south of the intersection of routes 340 and 343 light green to gray cherts are exposed with rare pyrite nodules less than 1 cm across.

In shoreline outcrops of the Luscombe Formation north of Campbellton manganiferous Member C and the uppermost Member G are clearly recognized. Member C in this area consists of gray chert with interbedded pink spessartite layers and a few jet black weathering rhodochrosite nodules and lenses.

In summary, the Luscombe Formation consists of 7 members in the vicinity of southern Loon Bay that gradually become less distinct westward towards Campbellton where the Luscombe Formation is more argillaceous and contains less manganese bearing minerals. In part, this may be due to the fact that the Campbellton section was measured along "dirty" but continuous woodland outcrop as previously discussed. This is perhaps also indicated by the fact that Members C and G can be clearly recognized in shoreline outcrops north of Campbellton.

Environment of Deposition of the Luscombe Formation

The Luscombe Formation consists mainly of argillaceous, manganiferous to non-manganiferous chert with minor interbedded limestones, volcaniclastics and (?) tuffs. The Luscombe Formation conformably grades upward from the Loon Harbour volcaniclastics that appears to represent the subsiding flank of a submarine fan deposit adjacent to a mafic/intermediate volcanic source. The existence of minor interbedded volcaniclastic and (?)tuff beds throughout the Luscombe Formation implies continuous nearby volcanism during the time of its deposition.

The manganiferous nature of the cherts within the Luscombe Formation is, of course, of critical importance to its mode of origin. Present-day manganese deposits are seen forming in marine environments as rounded nodules or as encrustations on volcanic rocks. Manganese nodules and crusts are found in a wide variety of environments from the deep-sea floors to fresh water lakes such as Lake Michigan, Wisconsin (Rossman and Callender, 1968) or Lake George, New York (Schoettle and Friedman, 1971; Morgenstein, 1972) to shallow marine areas such as the Kauai Channel in Hawaii (Morgenstein and Andrews, 1971). Bonatti and others (1972) further point out that tremendous variability occurs in the characteristics and geological settings in which ferromanganese deposits occur. These deposits are found in continental shelfs, deep-sea hemipelagic regions, pelagic abyssal plains, ocean ridges and seamounts.

Most manganese nodules occur at the sediment-water interface and develop in very deep water far removed from land where sedimentation rates are exceedingly slow. These nodules are generally associated with red clay and/or radiolarian chert deposits. Encrusting deposits may also be found on seamounts where sediment accumulation is negligible (Horn and others, 1972). Kidd and Armannson (1979) have recently described the occurrence of manganese and iron-rich micro-nodules with shallow water shell debris at a depth of 100 meters on the summit of

Mount Palinuro, a seamount off the west coast of Italy. Marine manganese nodules and encrustations are commonly found within active ocean ridge systems. Accretion of manganese deposits is not controlled by depth, pressure or temperature of the surrounding sea water since ferromanganese deposits are found in water as shallow as 5 meters in the Hawaiian Archipelago (Morgenstein, 1972).

Present-day deep sea marine red clays show a large excess of manganese with respect to the average amount of manganese found within granites, basalts and shales. These marine deposits are enriched in manganese by a factor of 7 higher than expected (see Broecker, 1974). As summed up by Broecker (1974) there are three major hypotheses that can account for the enrichment of manganese and other metals such as iron, nickel and copper in marine sediments. The first hypothesis is the separation of these elements during continental weathering. This idea contends that during the terrestrial weathering process the manganese and iron tend to attach themselves to very fine lightweight particles. This allows them to remain suspended for a long time to be spread quite uniformly over much of the ocean floor as detrital manganese and iron. This hypothesis does not seem to be applicable to the origin of the manganese within the Luscombe Formation because the manganese bearing minerals occur in discrete layers rather than being uniformly dispersed throughout the chert as would be expected if the manganese was detrital in origin.

The second major hypothesis involves secondary enrichment of surface deposits by the upward diffusion of iron and manganese through the underlying sediment column (see also McGeary and Damuth, 1973). A problem with this idea is that no manganese-depleted sediments have been found deep within the sediment column. Also, the accretion of manganese on basaltic bedrock in Hawaii does not support this theory (see Morgenstein, 1972).

The third important hypothesis suggests that manganese, iron, nickel and copper (along with silicon) are hydrothermally pumped into seawater in regions of subsea volcanism where these elements easily dissolve into hot sea water (Broecker, 1974). The most important areas of submarine volcanism are mid-ocean ridges and partially submerged island arcs and back-arc basins.

Lyle (1976) has argued that hydrothermal manganese introduced by volcanism along oceanic spreading centers is the major source of manganese input into the oceans. Minor amounts of manganese can also enter sea water by direct leaching of the newly emplaced basalt (Corliss, 1971). Because manganese is nearly insoluble in oxygenated sea water (Krauskopf, 1956), a small percentage of hydrothermally derived manganese as well as iron and SiO_2 is rapidly precipitated as Fe-Mn-SiO_2 crusts very near its volcanic source (Lyle, 1976; Toth, 1980). However, the majority of hydrothermally introduced Mn and Fe oxides and associated SiO_2 species remain in colloidal suspension to be carried away from their volcanic source area to eventually flocculate with other metals contained in sea water forming Mn-Fe-Si rich metalliferous sediment that are reworked by bottom currents (Toth, 1980). The Luscombe Formation was both spatially and temporally associated with intense submarine volcanism (for example, see Strong, 1977) making the volcanism theory most tenable for the origin of both the manganese and silica (chert), in the Luscombe Formation.

The origin of the chert within the Luscombe Formation must be intimately related to the origin of the manganese within this unit.

This chert could represent the diagenetic product of either colloidal precipitation of hydrothermally introduced silica from sea water, siliceous tuffs derived from a volcanic source, or the siliceous remains of sponge spicules and radiolarian tests. The existence of microscopic radiolarian-like spherulitic structures in the Luscombe cherts (less than 5% in most thin sections) suggests that at least a small portion of the chert is biogenic in origin. Similar radiolarian-like structures were found in the upper Ordovician Mount Merino cherts of the Taconic sequence of eastern New York (Lang, 1969) and in the middle Paleozoic Caballos Novaculite of the Marathon region, Texas (McBride and Thomson, 1970).

Petrographic studies of the novaculite in both the Arkansas and Caballos Formations lead Goldstein and Hendricks (1953) to conclude that the source of silica for these rocks was extensive volcanic ash falls deposited when little sediment was being supplied to the Ouachita 'geosyncline.' Submarine weathering removed the easily soluble elements from these tuffs converting them to nearly pure opaline silica. This allowed radiolarians and siliceous sponges to thrive in a high-silica environment with their remains also being preserved in the siliceous sediments. Conversely, Park (1961) and Park and Croneis (1969) argued for a purely biogenic rather than volcanogenic origin of the Caballos and Arkansas novaculites. Fan (1964) argued that the Caballos novaculite could be divided into both biogenic and volcanogenic units. McBride and Thomson (1970) claim that the Caballos Novaculite is of biogenic origin with the chert forming by the diagenetic alteration of opaline skeletal particles of sponge spicules and radiolaria. They further claim that the Caballos was deposited in a deep-marine trough adjacent to a peneplained land mass with a rate of accumulation of 0.1 to 0.5 mm/ 1000 years. Very rare manganese nodules are also found in the Caballos Novaculite (McBride and Folk, 1977). The regional geological setting and lithological character of the Luscombe cherts differs drastically from the Caballos Novaculites, therefore, no similarities in the mode of origin between these two units can be drawn.

Most of the present-day opal (SiO_2) deposits are forming in deep marine environments beneath enhanced upwelling zones such as the equatorial Pacific region where biological productivity is high (Broecker, 1974). It is unlikely that the Luscombe cherts originated beneath an upwelling zone because the ubiquitous non-penetrative soft sediment slump structures found in this unit (see Lithology section) imply a more unstable, tectonically active environment of deposition than a deep marine abyssal plain. Also, the numerous volcaniclastic and thin (?)tuff beds within the lower portion of the Luscombe Formation attest to the proximity of this unit to a volcanically active region during its early depositional history. The Luscombe Formation cannot represent abyssal marine sediments preserved on the flanks of a seamount subducted into the Dunnage melange subduction zone because the Dunnage melange does not represent the frontal part of a subduction/ accretion prism (discussed further in Conclusions section; see also Hibbard and Williams, 1979). Due to its regional geological setting, a volcanic origin for the Luscombe chert seems most probable. High amounts of colloidal and dissolved silica can be pumped into sea water by hydrothermal activity in regions of intense submarine volcanism (Lang, 1969; Toth, 1980). This silica will easily precipitate out of the sea water in regions surrounding the volcanic source (Taliaferro, 1934; Siever, 1962; Toth, 1980). Therefore, the Luscombe Formation must have been deposited in a marine environment proximal to a zone of subsea volcanism where hydrothermal systems pumped large amounts of both manganese and silicon into the surrounding sea water. Radiolarian may have thrived in this silica saturated water if enough nutrients were available.

The discontinuous layers of rhodochrosite and the irregularly-shaped pods of manganiferous limestone in Members C and F of the Luscombe Formation, respectively, represent once continuous manganiferous limestone layers. These former layers are so poorly preserved that it is not clear whether they represent deep or shallow water limestone deposits or exactly what bearing they have on the environment of deposition of the Luscombe Formation.

Riding Island Greywacke

The Riding Island Greywacke was first defined by Kay (1975) as the uppermost unit of the Campbellton sequence. The name was given because excellent outcrops of this greywacke are found on Riding Island and on the shores of Campbellton. In general, the Riding Island Greywacke consists of medium to very fine-grained, very thick bedded, Bouma-type sequences of quartz-bearing greywackes with associated mud tops. Inland, outcrops are quite numerous but due to the common very thick-bedded nature of the greywacke, bedding is difficult to discern in these outcrops. Good road cuts of this greywacke are seen west of Campbellton along route 340. Outcrops of the Riding Island greywacke are quite abundant but restricted to the region within and south of Campbellton. The present preserved thickness of the Riding Island greywacke is estimated to be about 1200 meters.

The lower contact of the Riding Island greywacke with the Luscombe Formation is nowhere exposed but can be inferred from structural relationships to be conformable. This opinion was also held by Kay (1975).

The Riding Island greywacke consists of rhythmically bedded, Bouma sequences of quartz and plagioclase rich, pebbly to sandy, graded greywackes with associated mud tops (Fig. 2-11). Greywacke portions of this unit are gray in color whereas the argillaceous portions are dark gray to green. Both argillite and greywacke weather gray with a rusty brown limonite to a reddish iron oxide stain. Bedding varies from 2 cm to 10 or more meters thick. Mud tops are sometimes finely laminated and less than 20 cm thick. Beds are commonly graded from pebbly (pebbles less than 1 cm across) at the base fining upward into very fine-grained sandy graywackes overlain by laminated argillite. Flame structures, load casts, and scour marks are extremely well developed and quite common (Fig. 2-12). Scour marks have up to 1 meter of relief. Some large intraformation scour-fill channels with several meters of relief are exposed along the northern shore of Riding Island proper. Trough cross-beds and cross-laminations ranging from centimeters to meters thick are also commonly found. Convolute soft-sediment fold structures are observed in places along the bases of the greywacke units.

The greywacke is very well indurated and always very poorly sorted with grain size varying from pevvles less than 1 cm across to very fine sand grains. The grains vary from rounded to angular and are usually equant in shape.

Figure 2-11: Rip-up clasts in Riding Island Greywacke.

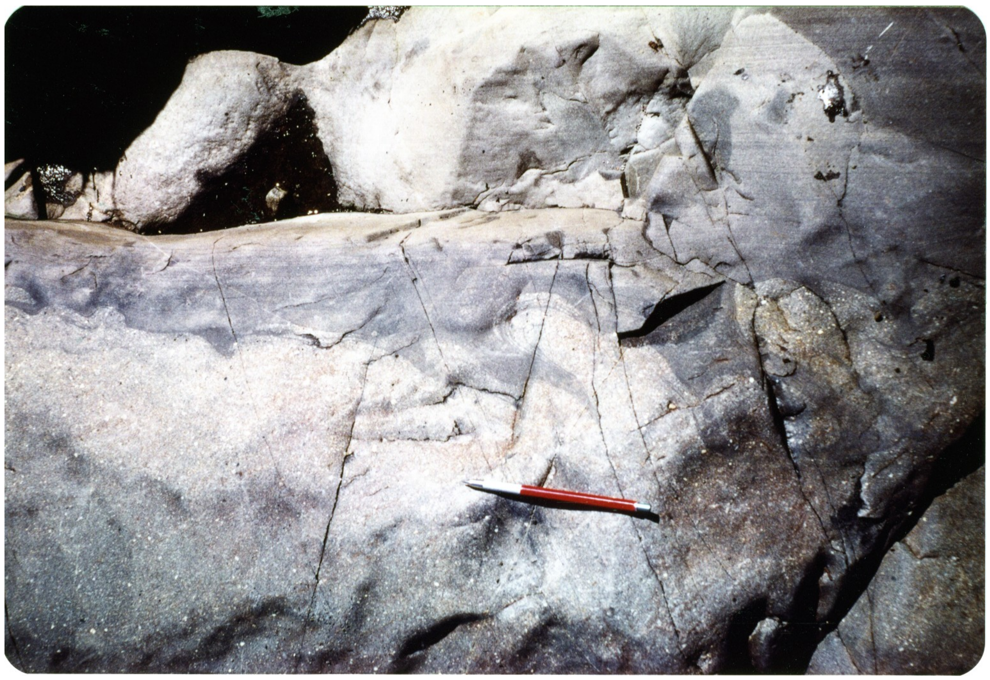


Figure 2-12. Bottom structures in Riding Island Greywacke.

Divisions A, B, D and E of the classical Bouma (1962) turbidite sequence are observed. Division E, of non-turbidite nepheloidal origin is rarely seen, but when present, consists of argillaceous chert. Mud tops exhibit parallel lamination characteristic of division D of the Bouma sequence.

A few thin micritic limestone interbeds are rarely found. Intra-formational angular rip-up clasts of laminated argillite or pebbly to fine-grained greywacke are found in both the greywacke beds as well as the mudtops constituting local intraformational breccias (Fig. 2-13). A large slump block of greywacke several meters wide was recognized along the shore just east of the Indian Arm Brook outlet.

Thin sections of the greywackes show that they are composed of approximately 20% each of quartz, chert or silicic volcanic and plagio-clase clasts, and less than 5% of each of orthoclase, microcline, silicic porphyry clasts, granitic to granodioritic clasts of plagioclase-quartz-microcline, fine-grained mafic volcanic clasts and micrographic granite clasts. Some of the quartz clasts exhibit resorption cavities suggesting they are volcanic in origin. The matrix of these greywackes occupies 20% of the rock and consists of very fine-grained recrystallized biotite with some chert and epidote. Thin sections show that the mud tops consist of 70% very fine-grained recrystallized biotite in optical continuity, 25% very fine clastic sand grains of plagioclase and quartz, and 5% opaques (pyrite?).

The origin of the unit is discussed in the stratigraphy portion of this thesis (see relationship between the New Bay Formation and Riding Island Greywacke section).

Figure 2-13. Rip-up clasts in Riding Island Greywacke.

New Bay Formation

The type section of the New Bay Formation was first defined by Helwig (1967) for outcrops along the east shore of South Arm in New Bay. Kranck (1952) had originally proposed the name New Bay for similar rocks but he did not specify a type locality. As defined by Helwig (1967) the New Bay Formation consists of a distinctive sequence of intercalated thick-bedded, polymictic, pebbly turbidites and thin bedded silty shales up to 1750 m (5800 ft.) thick. The New Bay Formation within my map area is found in the Michael's Harbour region and along Rideout Point north of Campbellton. Rocks along Rideout Point were formerly mapped by Kay (1975) as the Rideout Point Formation. Due to the lithological similarities between the rocks exposed along Rideout Point and the New Bay Formation in the Michaels Harbour area, I am compelled to drop the term Rideout Point Formation and include these rocks as part of the New Bay Formation. Maximum exposed thickness of the New Bay Formation in the Michael's Harbour area is approximately 1,000 m (3,300 ft.).

To the south of Michael's Harbour the New Bay Formation is in fault contact with the Botwood Group. Northward, westward and eastward from Michael's Harbour the New Bay Formation grades into the Dunnage melange. Lithology of the rocks does not vary greatly across this New Bay/Dunnage contact (discussed later).

The New Bay Formation consists of interbedded polymictic conglomeratic greywackes and thinly-bedded, fine-grained, sandy greywackes and argillites. The conglomeratic greywacke beds comprise up to 30 to 40% of the total outcrop and are mainly found along the northeastern arm of the Michael's Harbour peninsula along the shore just northeast of

Michael's Harbour proper, and along Rideout Point north of Campbellton. These conglomerate horizons are generally gray in color, weathering light gray with a reddish iron oxide stain. Bedding of these conglomerates varies from 1 m to 20 m thick in lenticular, laterally discontinuous beds that cut deep channels into the adjacent thin bedded portion of the New Bay Formation. In places, high angle truncation of the thin-bedded material against the conglomerate beds is seen. Bedding in the conglomeratic greywacke is always graded from conglomeratic pebbles and boulders at their bases finding upwards into fine-grained greywackes with associated mud tops. Scour marks and load casts with up to 1 m of relief are commonly found at the base of the conglomeratic greywacke beds. Mud tops in the fining upward beds rarely exhibit convolute bedding and flame structures.

Pebble content can vary from 10 to 70% of the total greywacke. The clasts are poorly sorted with sizes ranging from boulders up to 60 cm across to fine-grained pebbles with the average size being 3-5 cm across. The shapes of these pebbles in undeformed zones varies between equant and prolate and they are angular to subrounded. The largest boulders and pebbles are composed of limestone, mafic lava and fine- to medium-grained granodioritic clasts. Conglomerate clasts consist dominantly of detrital quartz, silicic porphyry, argillaceous chert, argillite, granodiorite, altered mafic lava (vesicular to fine-grained) and micritic limestone. Matrix of these conglomeratic greywackes is dark gray argillaceous material that is made of up to 50% fine to coarse sand size grains of angular to subrounded quartz, feldspar, silicic porphyry and other components of the pebbles floating in a much finer grained matrix of mainly biotite with minor amounts of sphene.

Angular rip-up clasts of the surrounding thin-bedded portion of New Bay Formation compose up to 30% of the conglomerates. These rip-up clasts are generally rectangular in shape and can be as large as 1 X 2 m across but average about 1 X 5 cm across. They are commonly imbricated parallel to bedding or are found in tight soft sediment folds. In places, pebbles are incorporated into the rip-up clasts. Numerous intraformational rip-up clasts of conglomeratic greywacke are also found. All of the sedimentary structures and features of these conglomeratic horizons imply that they are turbidite deposits (Helwig, 1967) that appear to be restricted to local channels cut into the thin bedded portion of the New Bay Formation.

The thin-bedded portion of the New Bay Formation is composed of rhythmically bedded, dark gray, fine-grained, sandy greywackes and silty argillites. Bedding is defined by alternating light and dark gray layers of greywacke and argillite, respectively. Bedding varies from laminated to thin bedded or 1 mm to 5 cm thick. Sedimentary structures are ubiquitous and include grading, convolute folding in zones less than 1 cm thick usually confined to individual beds, and non-penetrative soft sediment gravity faults with less than 1 cm of dip-slip. Sedimentary structures characteristic of zones B, C and D of the Bouma (1962) ideal turbidite can be identified in these rocks. The bases of the sandy, fine-grained greywacke beds exhibit minute load casts with 1 to 5 mm of relief and flame structures with up to 1 cm of relief. Above these basal structures the sandy, fine-grained greywacke is plane bedded (laminated; zone B of Bouma sequence) passing upward into climbing ripples (Zone C of Bouma Sequence). Only the thicker 3-5 cm beds contain both plane bedded and climbing ripple zones. Thinner 2-3 cm thick beds contain only climbing ripples. In these thinner beds, the climbing ripples occur in a continuous layer of eye-shaped packages less than 2-3 cm across or as disconnected eye-shaped climbing ripple packages resulting in flaser bedding. Above the plane bedded/climbing ripple zones is a plane bedded silty argillite layer (Zone D of Bouma Sequence) in beds up to 4 cm thick.

Thin sections of this thin bedded material shows that it consists of alternating zones of light colored, very fine-grained, volcaniclastic material and dark-colored, argillaceous material. The light-colored zones primarily contain very fine-grained, subhedral, broken plagioclase laths with subsidiary very fine-grained quartz(?), carbonate and chlorite. The dark-colored argillaceous zones are composed of a very fine-grained mixture of mica and plagioclase. In the Michael's Harbour and Emily Cove peninsulas the New Bay Formation passes westwards into a soft sediment disruption zone over a distance of several meters. In this zone manganese-bearing buff-colored layers exhibit what Hibbard (1976) referred to as 'toothpaste' structure. That is, these layers are chaotically folded and display wispy boundaries that commonly show fluid injection structures into the surrounding argillaceous material (Fig. 2-14). This zone is overprinted by a later cleavage that forms augen around the folded manganese-bearing buff-colored layers. Deformed conglomerate beds are found interbedded with this argillaceous material on the western portion of the Michael's Harbour peninsula.

Interbedded gray-green chert beds (?indurated tuff) 1-5 cm thick are rarely found in the thin-bedded portion of the New Bay Formation. Also interbedded in this thin bedded material are a few graded fine to coarse grained sandy greywackes in beds up to 4 meters thick and thin, micritic limestone lenses (nodularised beds) up to 2 X 20 cm across.

Found in both the conglomeratic greywacke and thin-bedded portions of the New Bay Formation are buff to dark brown weathering spherical limestone concretions up to 5 cm wide that weather out of the surrounding rock.

Interbedded with the conglomeratic greywackes in the Rideout Point area are recrystallized (hornfels facies from intrusion of the Loon Bay batholith) argillites that contain distinctive buff-colored layers (Fig. 2-14). These buff layers are highly disrupted and polyclinally folded. They are composed of mainly spessartite and quartz (described in more detail in Dunnage melange section). On the Emily Cove peninsula similar manganiferous buff layers appear to be developed within the thin-bedded portion of the New Bay Formation. Similar rocks are found along the shore in the Mussel Bed Rocks area where they form a highly cleaved matrix surrounding isolated blocks of argillaceous conglomeratic greywackes. Since these isolated blocks were found floating in an argillaceous matrix these rocks were mapped as Dunnage melange. Hibbard (1976) also mapped the Mussel Bed Rocks area as Dunnage melange for similar reasons. However, the argillaceous rocks that surround these greywacke blocks in the Mussel Bed Rocks area are very similar to the argillaceous zones associated with the conglomeratic greywacke near Rideout Point in that they both contain similarly deformed manganiferous buff-colored layers (the conglomeratic greywackes, however, do differ between these areas though as discussed in the Dunnage melange section). Since the conglomeratic greywackes in the Rideout Point area look more similar to the conglomeratic horizons found in the New Bay Formation in the Michael's Harbour area and since they do not appear to be structurally disrupted, forming a continuous zone parallel to the coast, this area was mapped as New Bay Formation.

Figure 2-14: Buff-colored layers in the New Bay Formation along Rideout Point.

The New Bay Formation has been established by Helwig (1967, 1969) to be of middle Ordovician (Llanvirnian/Arenigian) age. The turbidite beds of the New Bay Formation most likely were deposited on the flanks of the mafic to intermediate island arc system that developed from lower to middle Ordovician times in this area (for example, see Dewey and Bird, 1971 or Williams, 1979).

Dunnage Melange

The Dunnage Formation was first defined by Kay and Eldredge (1968) for outcrops in its type area around Dunnage Island in Dildo Run. Horne (1968) showed that outcrops of the Dunnage Melange could be traced southwestward from Dildo Run into the peninsula north of Lewisporte (Fig. 2-15). The term Dunnage 'Formation' has been supplanted by Dunnage 'melange' by later workers in this area (see Hibbard, 1976; Kay, 1976).

To date, the best information on the Dunnage melange is from Hibbard (1976) and Hibbard and Williams (1979) who mainly worked on the southwest portion of the Dunnage melange and from Kay (1972, 1976) who worked on the northeast portion of the Dunnage melange in the Dildo Run area.

The Dunnage melange in my map area is found along Mussel Bed Rocks and southward into the woodland area and along the eastern shores of Burnt Bay just west of Michael's Harbour. This area was also mapped by Hibbard (1976) and comprises a small part of the southeastern portion of the Dunnage Melange. Hibbard (1976) presented a complete detailed lithological description of the Dunnage melange in this area

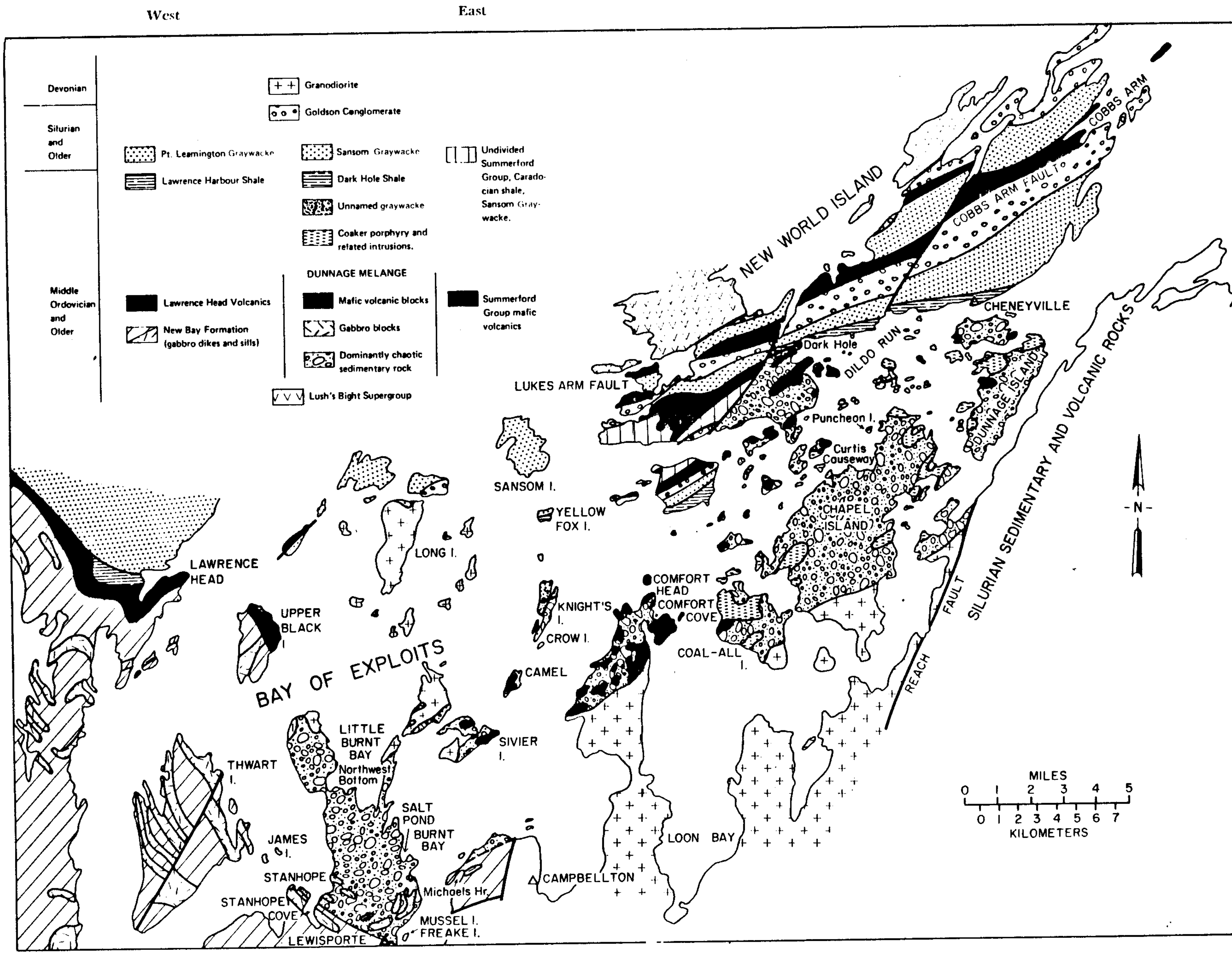


Figure 2-15 (from Hibbard and Williams, 1979) and no attempt is made here to supplant this data but simply to add to it. Since the Dunnage melange is quite variable in lithologic make-up it is not meant for this description to be extended to the whole of the Dunnage melange.

In the Mussel Bed Rocks area and southward into the woods the Dunnage melange consists of isolated blocks of pebbly greywacke floating in a matrix of dominantly argillaceous material. The greywacke blocks constitute about 30% of the total outcrop of the Dunnage melange in this area. Both the greywacke blocks and surrounding matrix material are of the same dark brown to dark gray color. The melange matrix material surrounding these blocks is composed of either argillite or argillaceous fine-grained, sandy graywackes. Portions of this matrix have been hornfelsed by the nearby Loon Bay batholith into highly recrystallized meta-semipelites that contain up to 10% medium-grained, cordierite porphyroblasts (on the eastern side of Mussel Bed Rocks this matrix material weathers with a sulfurous yellow to reddish iron oxide stain). Bedding is poorly defined to non-existent throughout the matrix portion of this melange. An important characteristic of the matrix material is that it contains from 5-20% thin, lenticular buff-colored layers that are a few cm (1-3) thick and between 2-20 cm long. Thin sections show that these buff layers are made primarily of very fine-grained euhedral spessartite crystals with lesser amounts of interstitial quartz and rare chlorite and (?)rhodonite. They are commonly zoned with an outer layer of 50% fine-grained spessertite and 50% interstitial recrystallized quartz surrounding an inner zone of predominantly fine-grained spessartite. On the Emily Cove peninsula these spessertite/quartz layers are mantled by a thin zone of fine-grained chlorite. Very fine-grained micaceous material is found surrounding the buff layers.

Thin sections of the meta-semipelites show them to consist of a very fine-grained matrix of quartz and biotite surrounding spherical cordierite porphyroblasts up to 1/2 mm across that are dominantly altered to very fine-grained pinite. Less altered cordierites exhibit characteristic sector twinning.

The greywacke blocks occur as either isolated single 'beds' several meters or less across surrounded by argillaceous matrix or as discrete packages containing several beds, usually of fining upward cycles of Bouma-type sequences that as a whole can be tens of meters across. Bedding in these greywacke packages strikes into and is truncated by the surrounding Dunnage melange matrix material implying that they are floating in this matrix.

Thin sections of some of the greywacke 'beds' show that they are composed of a very fine-grained mixture of recrystallized quartz, biotite, and chlorite with a few percent of finely disseminated fine-grained, euhedral spessartite crystals.

Bedding is recognized throughout the Dunnage melange along western Burnt Bay just west of Michael's Harbour. Even zones consisting of highly cleaved argillaceous matrix with intraformational pebbly mudstone melange clasts appear to be continuous and strike parallel with the well bedded portions of the Dunnage melange. These pebbly zones may represent discrete olistostrome horizons. Hibbard (1976) showed that intact bedded sections could be mapped within the Dunnage melange but did not map this portion of the Dunnage melange as such.

Middle Cambrian trilobite fragments have been recovered from a limestone block within the Dunnage melange at its type locality on

Dunnage Island (Kay and Eldredge, 1968). Tremadocian graptolites were found in the black shaly matrix material of the Dunnage melange near Stanhope (Hibbard, 1976). Arenigian conodonts were found in a limestone block along the New Bay Formation/Dunnage melange transition zone on James Island (Hibbard, 1976 as an addendum; Hibbard and others, 1977). The Dunnage melange is overlain by Caradocian black shales (Dark Hole shale) in the Dildo Run area (Horne, 1970). This suggests the Dunnage melange developed from lower to middle Ordovician times and is temporally correlative with the Middle Exploits Group (Lawrence Head Formation and New Bay Formation, Helwig, 1967). Hibbard (1976) and Hibbard and Williams (1979), claimed that the Dunnage melange interdigitates with and internally mimics the stratigraphy of the middle Exploits Group implying that the Dunnage melange is a lateral chaotic equivalent of the middle Exploits Group.

Manganiferous Nature of the Dunnage Melange

The thin disrupted, buff-colored layers found within the Dunnage melange exhibit a spessartite-quartz mineral assemblage due to thermal metamorphism in zones proximal to the comagmatic Loon Bay and Long Island batholiths (Fig. 2-16, see also Hibbard, 1976). According to Hibbard (1976) these layers are found throughout the argillaceous matrix material of the southwest portion of the Dunnage melange. Hibbard (1976) also reported that rounded to ellipsoidal garnet clusters (?nodules) are found in the matrix material of the Dunnage melange on Camel Island. Thin sections of both the hornfelsed buff-colored layers and the garnet-rich clusters of Camel Island by the author reveal that

Figure 2-16. Buff-colored layers in the Dunnage Melange on Camel Island.

these garnets are always distinctly anisotropic spessartite. The most manganese rich beds within the Dunnage melange are found on Camel Island (Hibbard, 1976).

A one to two meter thick unit of highly manganiferous chert is found on the northeast side of Camel Island that weathers with a bright metallic blue manganese oxide stain (Fig. 2-17). This unit looks somewhat similar to portions of member C of the Luscombe Formation and contains non-penetrative, soft-sediment folds. Thin sections of this bed show it to be composed of 70% very fine-grained, euhedral spessartite crystals and 30% microcrystalline quartz. Just north of this bed on Camel Island is a zone containing numerous flattened manganese-rich nodules set in an argillaceous matrix (Fig. 2-18; see also plate 1 of Hibbard and Williams, 1979). These nodules are ellipsoidal in shape and up to 6 cm across. A penetrative 'hard-rock' cleavage forms augen around these nodules (Hibbard, 1976). The longest axes of the nodules is locally parallel with this cleavage, but not consistently. This suggests that the flat shape of the nodules is not a result of flattening into this cleavage. Rather, this could be the original shape of these objects or they could have been flattened during a period of soft-rock deformation. Manganese-bearing buff-colored layers were found in the New Bay Formation on Michael's Harbour near the Dunnage melange contact. Hibbard reports finding similar disrupted buff-colored layers on Thwart and James Islands near the contact with the Dunnage melange. These layers are similar in appearance to the spessartite bearing layers found in both the Dunnage melange and New Bay Formation adjacent to the Loon Bay and Long Island batholiths. If the buff-colored layers found in other portions of the New Bay Formation

Figure 2-14: Manganiferous chert bed in the Dunnage melange on Camel Island.

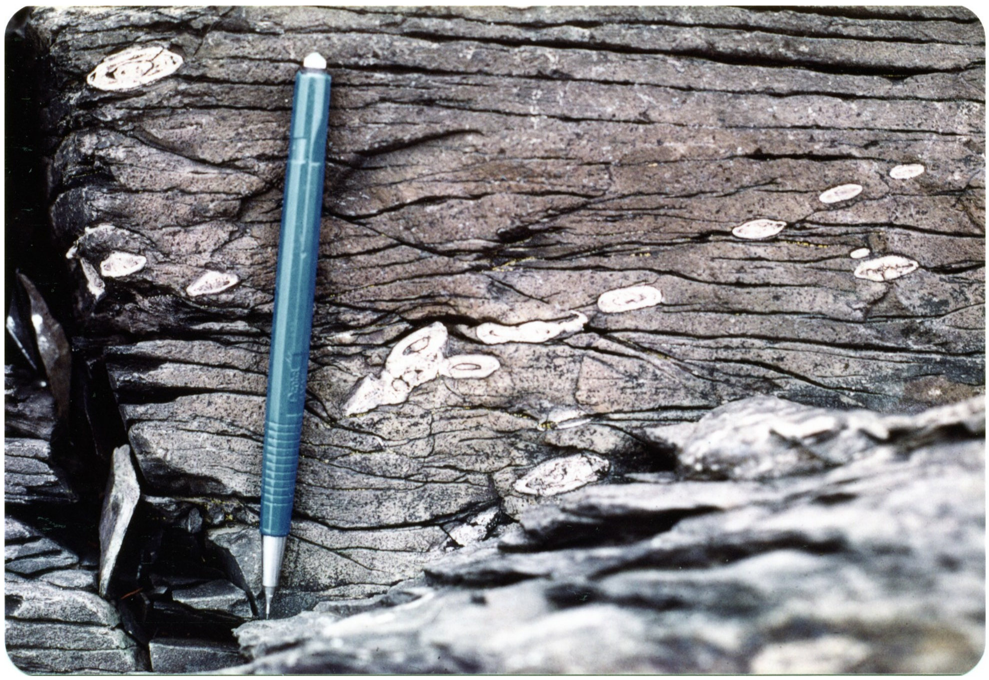


Figure 2-18. Manganese rich nodules in argillaceous matrix of the Dunnage Melange on Camel Island.

adjacent to the Dunnage melange such as on Thwart and James Island were also thermally metamorphosed they would certainly yield manganese garnet porphyroblasts. Since the Dunnage melange was derived from the New Bay Formation (Hibbard, 1976), it logically follows that both units should contain manganese-bearing layers. It is not clear if just the New Bay Formation proximal to the Dunnage melange is manganese-bearing or if all of the New Bay Formation throughout the type section in New Bay proper (see Helwig, 1967) is manganiferous. As an intuitive guess, I would say that most of the thin-bedded portion of the New Bay Formation is somewhat manganiferous since these rocks probably had the same provenance. However, this is not obvious from direct field observations (W.S.F. Kidd, pers. comm.).

The fact that both the Dunnage melange, New Bay Formation and adjacent Luscombe Formation are manganiferous cannot be fortuitous. Rather, the manganese in both these units is most likely related to volcanism (see Environment of Deposition of Luscombe Formation section). As discussed, the Luscombe Formation appears to have been deposited by the direct colloidal precipitation of volcanogenic silica and manganese from hydrothermal subsea vents whereas manganese in the Dunnage melange and the New Bay Formation came from the subaerial weathering of perhaps the same manganese bearing volcanics. Published major element chemical analysis of mafic Ordovician volcanic rocks from central Newfoundland (Kean and Strong, 1975; Strong, 1977) do not show any abnormally high amounts of manganese. Rather, they contain between 0.10 and 0.80 weight percent manganese which is normal for mafic volcanic rocks. Perhaps most of the highly manganiferous volcanic rocks were simply weathered away during Ordovician times to provide the New Bay Formation with volcaniclastic detritus or that somehow manganese was preferentially concentrated into these sediments by weathering processes.

Origin of the Dunnage Melange

Evidence found within my map area further supports the general notion of Hibbard (1976) and Hibbard and Williams (1979) that the Dunnage melange represents the lateral chaotic equivalent of the middle Exploits Group (New Bay Formation and Lawrence Head volcanics). However, they further state that the Dunnage melange formed in one single slump event synchronous with the intrusion of associated plutons such as the Coaker porphyry and that this slump event was the forerunner of a major change in tectonic regimes in central Newfoundland from active arc volcanism to arc degradation and basin infilling. Several features given by Hibbard himself imply, however, that the Dunnage melange actually formed by a continuous process of small intermittent slump events. The first line of evidence is the existence of intraformational pebbly mudstone rip-up clasts (that are internally chaotic themselves) within the Dunnage melange (see plate 6 of Hibbard and Williams, 1979). These clasts suggest a history (as given by Hibbard and Williams, 1979) of deposition of pebbles into mud and subsequent slumping, followed by semi-lithification and a second slumping event to disrupt the bed and re-incorporate pebbly mudstone clasts into beds of the same material. Hibbard and Williams (1979) go on to say that similar multi-stage soft-rock histories of deposition, ripping-up or slumping, redeposition and reslumping or ripping-up can be inferred from numerous blocks of sedimentary rocks found within the Dunnage melange. In addition to this, Hibbard (1976) reports finding pre-cleaved conglomeratic shale blocks in an argillaceous melange matrix on the northwest portion of the Comfort Cove peninsula. The regional cleavage that affects the surrounding rocks is oblique to the cleavage in the blocks, and therefore it is not the same. Hibbard (1976) claimed that the cleavage within the blocks may be an earlier cleavage that could be explained by a mechanism of soft sediment flow resulting in the development of divisional planes characterized by an alignment of material. This further implies that a series of complex slump and re-slump events occurred over a long period of time.

Numerous local well bedded (non-disrupted) zones found in the Dunnage melange (see Hibbard, 1976) could represent laterally continuous horizons deposited between periods of larger scale slumping events. Since the Dunnage melange is found outcropping on numerous widely scattered islands and shorelines around the Bay of Exploits and Dildo Run and since it has been modified by later plutonism, folding and faulting events, lateral continuity of these beds cannot be proved or disproved. In summary, evidence given above strongly suggest that Dunnage melange formed over a relatively long period of time by a series of intermittent slump events induced by gravitational instability into a filling basin and cannot have resulted from one single slump event.

Burnt Bay Chert (new)

This unit is distinguished by differences in its lithological character, and consists of bioturbated greey to gray chert and quartz-rich conglomerates. This unit underlies a large portion of the western part of the mapped area. A few good road cuts as well as a large quarry of Burnt Bay chert are exposed along route 340 south of Michaels Harbour. Inland, outcrops are exposed mainly along two forest access roads where separate outcrops of this unit are good but generally widely spaced.

The thickness of this unit is difficult to ascertain due to suspected large scale internal folding of the unit. As a guess, this unit is at least ± 500 meters thick. Unfortunately, none of the contacts with the surrounding units are exposed. Field relationships suggest that the eastern and western margins of this unit are in fault contact with both the Silurian Goldson conglomerate and Botwood Group. This unit appears to be in contact with the Riding Island greywacke along its northeastern margin. Here, field relations are equivocal; but my opinion is that this is a conformable stratigraphic contact with the Riding Island Greywacke passing upwards into the Burnt Bay Chert. However, the possibility that this is also a faulted contact does exist.

The Burnt Bay Chert consists of interbedded greenish bioturbated cherts, conglomerate, and fine-grained greywacke with minor amounts of argillite and carbonate. The cherts constitute the major portion of the unit. They are gray to green, intensely bioturbated, tuffaceous looking cherts that weather brown to white (Fig. 2-19). Bioturbation burrows on the fresh surface are darker in color than the surrounding chert. The burrows are less than 1 cm wide and up to 5 cm long. They are elongate both parallel and perpendicular to bedding.

The chert is laminated to thin bedded (less than 1 mm to 9 cm thick beds) with bedding well defined by green-grey color banding. Zones rich in very fine-grained sand-size clasts of plagioclase are rarely seen in the chert.

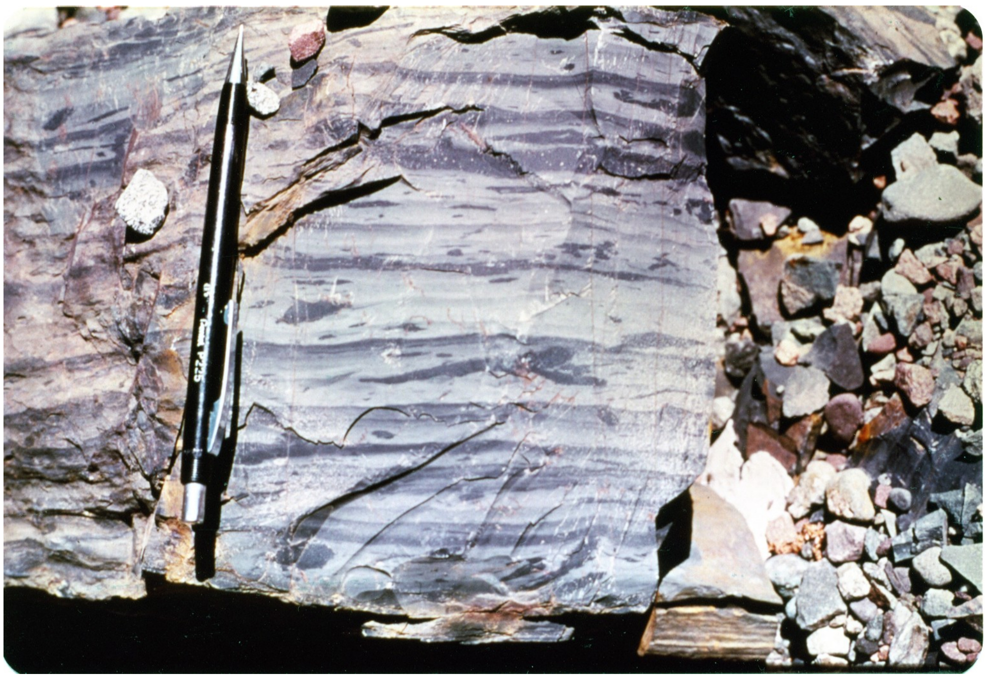


Figure 2-19. Bioturbation in Burnt Bay chert.

Surficial bluish metallic manganese oxide stains are seen in some sections of the chert. Thin sections show that this bluish strained chert consists of subequal amounts of crystalline carbonate (rhodochrosite) and quartz with up to 20% opaque manganese oxide stain. Thin sections of the gray-green chert show it to be composed of very fine-grained crystalline quartz with up to 15% very fine grains of chlorite and a few percent of detrital plagioclase. In thin section the darker bioturbation blebs are no different than the surrounding chert except for the increased presence of very fine-grained disseminated opaque material.

Thin interbedded buff-colored micritic carbonate layers are sometimes found in the chert. These carbonate laminations are less than 1 cm thick and are spaced every few centimeters in the chert in some outcrops. Soft sediment convolute laminations are sometimes exhibited by these carbonate layers.

Interbedded with the chert are numerous graded conglomeratic to fine-grained sandy greywacke beds. Bedding in these greywacke and conglomerate beds is between 1-3 meters thick. Scour marks, load casts and flame structures are common at the base of the beds. The beds usually fine upward into laminated mud tops with convolute laminations and vertical worm burrows. Pebbly zones consist of up to 85% angular to rounded equant to ellipsoidal pebbles less than 1 cm across (Fig. 2-20). These poorly-sorted conglomeratic zones are clast supported consisting of pebbles that are sometimes imbricated. Pebbles are set in a gray-green, quartz and plagioclase-rich, coarse to fine-grained sandy greywacke matrix. Thin sections show the pebbles to consist of microcrystalline chert or silicic volcanic clasts, quartz (appears

Figure 2-20. Conglomerate beds in Burnt Bay chert.

bluish in hand specimen), plagioclase (sometimes zoned), mafic lava, granophyre, silicic porphyry and dioritic to granodioritic clasts set in a matrix of coarse to fine-grained greywacke containing sand size clasts of the same lithology as the pebbles in a framework of fine-grained recrystallized biotite. Numerous large angular rip-up clasts up to 10 by 50 cm across are found in these pebbly zones. They are commonly imbricated and consist of mainly green to gray chert. Uncommon angular rip-up clasts of gray limestone and dark friable argillite are also found.

A sedimentary dike of fine-grained greywacke 20 cm thick perpendicular to bedding was found at one locality.

Southeast of the quarry along route 340 a forest access road trends southward through outcrops of mainly bioturbated green cherts. A gray porphyritic gabbro stock is found intruding the Burnt Bay chert in this area. Along the contact zone with this stock the green cherts have been hornfelsed and exhibit a deep maroon color. Farther south along this forest access road (south of the sawmill) outcrops of maroon chert alternate with outcrops of porphyritic gabbro. The southernmost outcrops consist of very argillaceous chert that exhibits a metallic blue manganese oxide stain interbedded with medium to fine-grained sandy greywackes. Scour marks and load casts at the base of some of the greywacke beds suggest that this section faces northward implying that these clastic rocks stratigraphically underlie the green bioturbated cherts. These rocks are very similar to portions of the New Bay Formation implying that the Burnt Bay chert stratigraphically overlies the New Bay Formation (see Stratigraphic Relationships section).

The green bioturbated cherts of this unit may represent the diagenetic alteration of volcanic ash fall beds. The small amounts of manganese oxides found in this unit further imply a volcanogenic origin of this unit (see discussion on manganese in environment of deposition of Luscombe Formation). No evidence was seen in thin section such as the remains of Radiolaria or sponge spicules to suggest a biogenic origin for this chert.

Unnamed Caradocian Black Slate

Good outcrops of this black graptolite-bearing slate are found on Shoal Point along the eastern shore of Burnt Bay just across the bay from Lewisporte.

Due to the highly deformed character of this unit, its thickness is quite difficult to estimate. From regional considerations of similar Caradocian-age black slates in the Notre Dame Bay area. Dean (1978) claims that the thickness of this slate in the Burnt Bay area is approximately 300 meters.

Due to the highly disrupted state of this unit and its limited outcrop extent the nature of the contacts with surrounding units is difficult to discern. Structural relationships suggest that the north and south contacts are fault contacts with the Wigwam sandstone and Goldson conglomerate, respectively. Regional stratigraphic relationships, as discussed elsewhere, imply that the unnamed Caradocian black slate conformably(?) overlies the Burnt Bay chert.

This unit is composed dominantly of friable, pyrite bearing, laminated, graptolite-bearing black slate that weathers light gray to olive green with sulfur yellow and rusty brown stains. Rounded pyrite nodules 1-3 mm across are commonly found in this slate.

Rarely interbedded with the black slate are greenish-gray microcrystalline chert beds generally less than 1 meter thick.

Graptolites collected by Williams (1964) have been identified as *Dicranograptus ramosus longicaulis* and *Climacograptus* corresponding to middle to upper Caradocian (Dean, 1978).

Goldson Formation

The Goldson Formation was first defined by Twenhofel and Shrock (1937) for outcrops along the northern part of New World Island. The Goldson Formation consists of dominantly coarse pebbly conglomerates with minor interbedded fine-grained, sandy greywackes.

The Goldson Formation is found throughout the southern part of the mapped area. Outcrops along the southeast and south-central portion of the mapped region are sparse woodland outcrops whereas good shoreline outcrops of this formation are found along the southeast shores of Burnt Bay. The thickness of the Goldson Formation in the mapped area is at least 2500 m (8300 ft). The Goldson Formation is in fault contact with all units to the north of its outcrop area.

Conglomerates of the Goldson Formation are dark gray to dirty green in color. Bedding varies from 20 cm to 7 m thick and is defined by fining upward cycles. Pebbles within the conglomerate commonly fine upwards and rarely are found to fine downwards (reverse grading). Flame structures, scour marks and load casts with up to 50 cm of relief are common. Soft-sediment slump structures are also frequently seen. Pebbles are rarely imbricated. Most conglomerates are clast-supported but the upper portions of some fining upward beds are matrix supported. Pebbles are poorly sorted, angular to rounded, and can be as large as 15 cm across but average about 1 cm.

Pebbles compose up to 80% of the total rock and consist of chiefly angular to subrounded clasts of silicic porphyry, very fine-grained mafic lava, granodiorite that rarely exhibits granophyric texture, cleaved argillite, fine-grained angular quartz-rich sandstones, and chert. Abundant fine to coarse-grained quartz and feldspar sand is found in the interstices between larger clasts. The plutonic granodioritic pebbles tend to be the largest. Most pebbles have an equant to prolate shape. Intraformational angular rip-up clasts of greywacke are common and can be as large as 20 cm by 1 m across but average about 10 by 20 cm.

Matrix of the conglomerate varies from gray to green argillite to fine- to medium-grained sandy greywacke containing sand grains of the same components as the conglomerate itself.

The subordinate non-conglomeratic portion of the Goldson Formation consists of fine-grained sandy greywacke that is dark grey when fresh, weathering to rusty brown with a reddish iron oxide stain. Bedding ranges from thinly laminated to 2 meters thick where it is defined by fining upward cycles. Flame structures, load cast, climbing ripples and flaser bedding are common in these graded greywackes. These rocks consist of 20% well sorted, rounded, fine-grained quartz sand grains, 5% rock fragments set in a silty argillaceous matrix that composes 75% of the rock.

Faunas in the Goldson Formation are of Llandoverian age where collected in shoreline outcrops along southeastern Burnt Bay (Williams, 1963). Faunas of early Silurian (Llandoverian) age have also been collected from the Goldson Formation in localities outside the Lewisporte-Loon Bay area (Williams, 1963; Kay, 1967).

Botwood Group

The Botwood Group as redefined by Williams (1963, 1972) consists of the lower Lawrenceton Formation and the upper Wigwam Formation.

The contact between these two formations is alleged to be conformable (Williams, 1962, 1967; Kay, 1969). Shelly faunas of Late Llandoverian-Early Wenlockian age have been collected from the Wigwam Formation in the Bay of Exploits area (Twenhofel and Shrock, 1937; Williams, 1967; summarized by Dean, 1978). Llandoverian corals were collected from a conglomerate bed of probable equivalence to the Lawrenceton Formation on the Change Islands by Eastler (1969).

Lawrenceton Formation

The Lawrenceton Formation consists of intermediate to silicic volcanic rocks. This formation is exposed in road cuts along route 340 just east and south of Michaels Harbour and along the shores of Rice and Freak Islands in Burnt Bay. Outcrops are also found along route 342 just north of Lewisporte. Exposed thickness of the Lawrenceton Formation in the mapped area is approximately 400 meters (1200 ft.). The Lawrenceton and overlying Wigwam Formations are everywhere in fault contact with the surrounding units.

The Lawrenceton Formation in the map area consists of gray-green, very fine-grained, silicic lavas that weather to a light brown color. Bedding is defined by light to dark gray-green color banding as well as flow banding. The tuffs have a homogeneous green matrix with small irregular dark gray zones (disseminated chlorite) less than 3 mm across. Within these tuffs angular agglomerate beds with breccia pieces up to 12 cm across are found. These agglomerate zones are up to 1 meter thick. Lava beds containing up to 50% subhedral plagioclase laths, with trachytic texture less than 3 mm large are common. Vesicles and carbonate-filled amygdales are also widespread within these tuffs. Limestone inclusions up to 6 cm across are rarely found in this unit.

Thin sections show that these tuffs are composed of very fine-grained subhedral plagioclase laths with trachytic texture set in an unidentifiable anisotropic matrix. Irregular blebs of chlorite less than 3 mm across occur as either vesicle fillings or replacing former mafic phenocrysts.

Wigwam Formation

The Wigwam Formation is composed of gray, coarse to fine-grained, silty, micaceous, quartz-rich sandstone that weathers brown-buff to red. The sandstone is thin to thick bedded in beds between 1 cm and 50 cm thick. Bedding is defined by faint color bands. The sandstone is either plane bedded or cross-bedded in wedge to trough-shaped cross-bed sets up to 10 cm thick. Well developed asymmetric ripple marks with wavelengths of 10 cm or less, and mudcracks, are found within the plane-bedded portions of the sandstone.

Thin sections of the sandstone show it to be composed of 60% very fine to fine-grained, angular to subangular, moderately sorted, quartz grains; 30% interstitial calcite and 10% very fine-grained secondary chlorite.

Sedimentary features of the Wigwam Formation such as cross-bedding, ripple marks, and mudcracks strongly suggest that this unit is a terrestrial fluvial deposit.

Intrusive Rocks

Mafic Dikes, Sills and Stocks (map unit A)

Numerous mafic dikes, sills and stocks are found intruding several units throughout the mapped area.

Dark greenish-gray gabbroic dikes and sills are commonly found intruding the New Bay Formation. These intrusions are always porphyritic (plagioclase) with the dikes and sills ranging up to 20 to 30 meters thick. Hibbard (1976) reported that the small island just north of Michael's Harbour was entirely made of porphyritic gabbro, perhaps representing a small stock. Contacts with the surrounding country rocks are always intrusive when seen. Grain size in these intrusions commonly increases inward from fine-grained chill margins to coarse-grained porphyry at the center of the intrusions. Thin sections of these gabbros show them to be composed of 50% coarse to fine-grained, sericitized, subhedral plagioclase phenocrysts; 40% fine-grained matrix composed of intergrown actinolite, chlorite and sphene; 5% of very fine-grained interstitial quartz; and 5% opaque magnetite.

Several east-west trending gray-green gabbroic dikes ranging from 4 to 25 m wide intrude the Riding Island greywacke. Thin sections of these dikes show that they are made of 60% medium- to fine-grained, subhedral, sericitized plagioclase phenocrysts that are sometimes broken with chlorite beards between the broken grains; 30% anhedral secondary actinolite and brown hornblende that mantles or completely alters augite phenocrysts; 10% fine-grained mixture of chlorite, epidote, sphene, calcite and opaque magnetite.

A well exposed small stock of gray porphyritic gabbro is found intruding the Luscombe Formation in the center of a woodland clearing just south of the intersection of route 340 with the Comfort Cove turn-off (route 341). This stock is about 150 meters wide forming a roughly circular outcrop pattern. Thin sections of this gabbro show it to be composed of 65% coarse to fine-grained sericitized subhedral plagioclase phenocrysts; 20% secondary actinolite and green-yellow hornblende; and 10% fine-grained epidote and chlorite.

Intruding the Riding Island greywacke and Luscombe Formation are several distinctive looking pale green porphyritic dikes ranging from 1 to 3 m thick. Thin sections of these dikes reveal that they are composed mainly of fine- to coarse-grained phenocrysts of plagioclase that is partially sericitized, surrounded by an extremely fine-grained green-yellow matrix possibly made of a mixture of chlorite and other material. These dikes may be similar to the greenish gray dikes discussed above.

A gray to green porphyritic gabbroic stock up to 3 kilometers across is found intruding the Burnt Bay chert. Numerous gabbroic dikes and sills are also found intruding this unit. Thin sections show this gabbro to consist of 50% subhedral, fine to medium grained plagioclase phenocrysts that are partially sericitized. The remaining 50% is composed of a fine grained groundmass of epidote, plagioclase, chlorite and actinolite.

All of the dikes, sills and stocks that intrude the New Bay Formation, Riding Island greywacke, Luscombe Formation and Burnt Bay chert are petrographically quite similar in appearance and are most likely comagmatic. Hibbard (1976) recognized that such gabbroic intrusions are very common and localized within the Dunnage melange and nearby New Bay Formation. According to Hibbard (1976) these intrusions were emplaced coevally with the formation of the Dunnage melange. All of the intrusions are also deformed with the surrounding rocks, implying that they are pre-kinematic to the regional, steep cleavage and folding.

Loon Bay Batholith

The Loon Bay batholith extends northward from the Michael's Harbour/Loon Harbour area towards Comfort Cove, encompassing most of the Comfort Cove peninsula and all of the Birchy Bay North peninsula (see Williams, 1963). Strong and Dickson (1978) claim that this batholith is an undeformed, post-tectonic intrusion into Ordovician sediments. However, along the western part of Mussel Bed Rocks where the Loon Bay batholith is seen intruding the Dunnage melange, cleavage in the Dunnage cuts through rocks of the Loon Bay batholith. Therefore, this intrusion may not simply be a post-kinematic (post-tectonic) intrusion.

The Loon Bay batholith is composed of white, coarse-grained, hypidiomorphic granodiorite. Thin sections show it to consist of mainly euhedral zoned plagioclase, euhedral orthoclase, anhedral quartz, biotite altering to chlorite, and primary hornblende altering to actinolite with minor amounts of sphene and apatite.

An aplitic, white, granodiorite dike from the Loon Bay batholith was seen intruding the Luscombe Formation along the northeastern shore of Loon Harbour. This dike is 1 meter thick and trends north-south. No other dikes were found.

Mesozoic Age Lamprophyre Dikes

Intruding all the units in the map area are numerous brown-colored dikes generally less than 1 meter thick. K/Ar age dates on similar intrusions to the west give early Cretaceous ages for these rocks (Helwig and others, 1974). Williams (1963) mapped a large stock of ultramafic and mafic plutonic rocks just south of Dildo Pond that is likely to be the source of these dikes.

These dikes are readily distinguished in the field since they are undeformed and cut across all previous structures. They have a distinctive brown color and contain numerous coarse-grained euhedral pyroxene phenocrysts. Thin sections of these dikes show that they are composed of 50% euhedral augite phenocrysts up to 1 cm across set in a very fine-grained, brown groundmass that is made of very fine-grained opaque magnetite, secondary calcite, fine-grained subhedral plagioclase, and fine-grained augite.

CHAPTER III

STRATIGRAPHIC RELATIONSHIPS

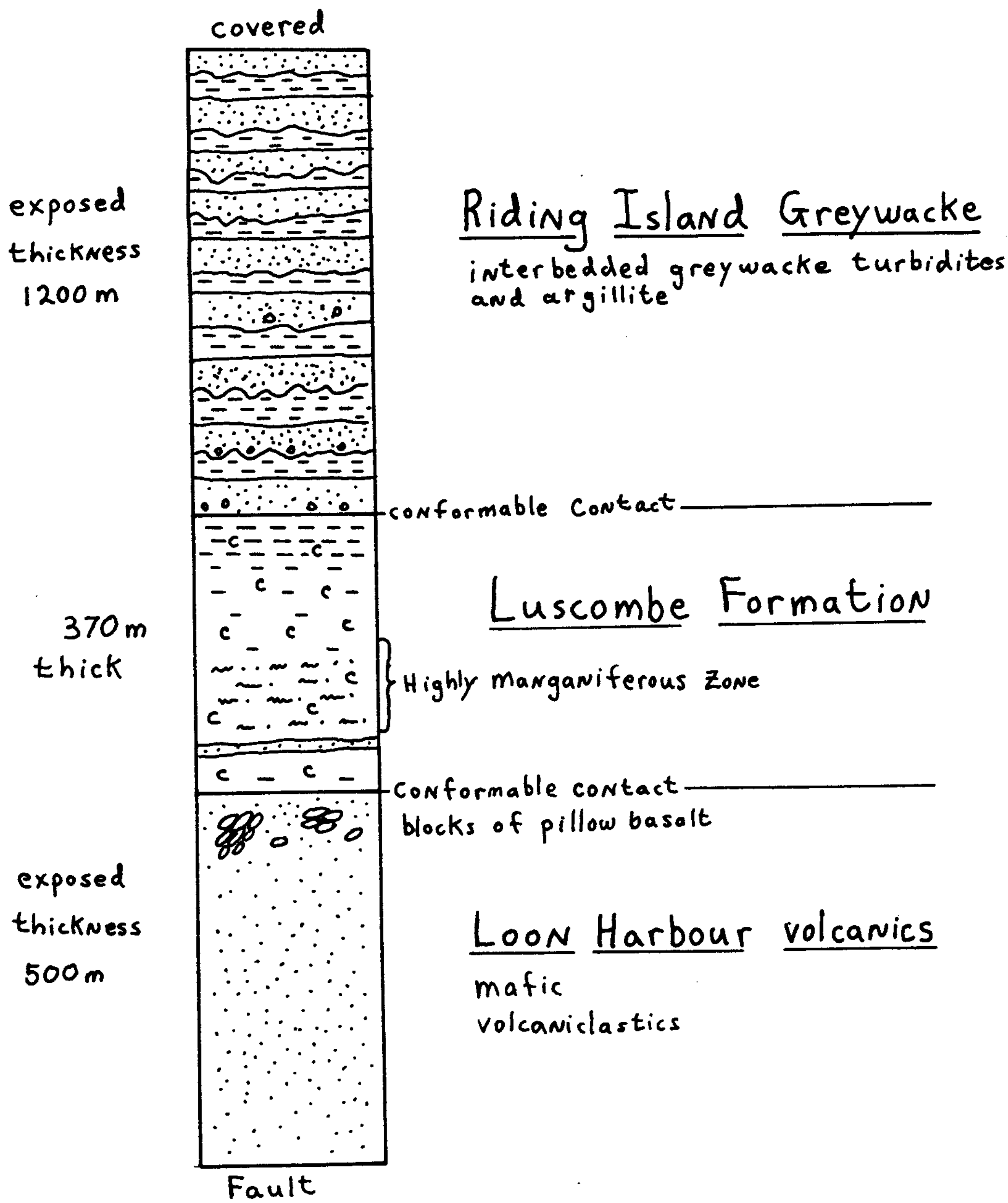
The Campbellton Sequence

As defined by Kay (1975) the Campbellton sequence consists of the Loon Harbour volcanics, Luscombe Formation and Riding Island greywacke. The fact that the Loon Harbour volcanics grades upward into and is conformably overlain by the Luscombe Formation is indisputable since the contact is well exposed and good facing evidence is found in both units. The Luscombe Formation is then overlain by the Riding Island greywacke as further demonstrated by facing evidence, but the actual contact is not exposed. The contact between these units is always separated by zone of no outcrop that is several meters wide. Along the shoreline in the northernmost part of Campbellton where good outcrops of both these units are exposed a relatively undeformed well bedded section of Riding Island greywacke passes northward into an equally well bedded section of Luscombe Formation with only a narrow zone of 2 to 3 meters of no outcrop between both of the units. Since the attitude of bedding does not change across this contact it is inferred that this is a conformable contact. A generalized stratigraphic column of the Campbellton sequence is illustrated in figure 3-1.

Age of the Campbellton Sequence

Portions of the Campbellton sequence are intruded by both the Loon Bay batholith and a series of very similar porphyritic gabbroic dikes, sills and stocks. Intrusive contacts of the Loon Bay batholith

Schematic Stratigraphic Column
of the Campbellton Sequence



Total thickness = 2070 meters

Figure 3-1

with the Luscombe Formation and Riding Island greywacke are well exposed. Since this batholith intrudes the Campbellton sequence isotopic age dates of the Loon Bay batholith should give a minimum age for the Campbellton sequence. Unfortunately age dates of this intrusion are conflicting and most likely unreliable. Wanless and others (1964) reported a K/Ar age of 450 ±20 m.y. B.P. for biotites from this batholith. Kay (1976) reported K/Ar age dates of 372 ±10 and 365 ±10 m.y. B.P. for biotites from recrystallized xenoliths found within the Loon Bay batholith. Several kilometers northwest of the Loon Bay batholith is the similar looking Long Island batholith that Heyl (1935) suggested was cogenetic with the Loon Bay batholith. Strong and Dickson (1978) presented petrographic and major element geochemical evidence supporting Heyl's (1935) assumption. K/Ar dates on biotites from the Long Island batholith gave an age of 440 m.y. B.P. (M. Kay, pers. comm., 1973, cited in Strong and Dickson, 1978). The best that can be said for this conflicting data is that it gives a minimum age for the emplacement of these batholiths implying that the Campbellton sequence is pre-365 m.y. B.P. (pre-Devonian) or more likely pre-450 m.y. B.P. (pre-Silurian) in age.

More definitive evidence for the age of the Campbellton sequence comes from the fact that both the Riding Island greywacke and the Luscombe Formation are intruded by dikes, sills and stocks of porphyritic gabbro that are petrographically similar to the gabbro intrusions that invade the nearby New Bay Formation and Dunnage melange (see Intrusive Rocks section). Since mafic volcanism ceased after Caradocian times in central Newfoundland (Nelson and Casey, 1978; Nelson, 1979) these intrusions are most likely pre-Caradocian in age. Hibbard (1976) claimed that these intrusions acted as feeder dikes for the pre-Caradocian

Lawrence Head volcanics found just to the east of the Lewisporte/Loon Bay area. Helwig (1967, 1969) maintained that the Lawrence Head volcanics are of Llandeilian age. If the gabbroic dikes that intrude the Riding Island greywacke are related to this magmatic event then the Campbellton sequence must be older than middle Ordovician. Kay's (1975) intuitive assumption that the Campbellton sequence is of upper Cambrian to early Ordovician age may be correct.

Age of the Burnt Bay Chert

The unnamed Caradocian graptolite-bearing black shale unit is found on Shoal Point in Burnt Bay adjacent to the distinctive green bioturbated cherts that typify the Burnt Bay chert. The contact between the two units is not exposed here. However, sequences of Caradocian black shale overlying green bioturbated chert are found in many areas of central Newfoundland. Nelson (1979) reported a similar sequence in the Seal Bay-Badger Bay area in what Dean (1978) termed the Shoal Arm Formation. Within the Shoal Arm Formation a green-gray bioturbated chert unit with a local metallic blue manganiferous stain passes upwards into black graptolite-bearing slate of Caradocian age. The green bioturbated cherts of the Shoal Arm Formation are identical in every way to the Burnt Bay cherts. Another similar section is found overlying the Dunnage melange in the Dildo Run area. Here the upper part of the Cheneyville Formation is composed of a green bioturbated chert unit 50 meters thick (Kay, 1970) that is similar to the Burnt Bay chert. This unit is overlain by the Dark Hole shale of Caradocian age. Both the Cheneyville Formation and the Burnt Bay chert contain conglomerate beds. Also, a block of bedded green-gray bioturbated chert with a blue manganese oxide stain was found floating in the deformed argillaceous matrix of the Dunnage melange in the Dildo Run area (W.S.F. Kidd, pers. comm.) that is similar in appearance to the Burnt Bay chert. Another such sequence is found in road cuts along route 35 south of Point Leamington. Here black shales containing Caradocian age graptolites overlie green bioturbated chert that resemble the Burnt Bay chert, except that it does not contain any manganese oxide stains. Extrapolating these regional stratigraphic considerations to the Burnt Bay area implies that the Caradocian black shale (?)conformably overlies the Burnt Bay chert. This indicates that the Burnt Bay chert is most likely of Llandeilian age. The fact that the Burnt Bay chert is intruded by porphyritic gabbro stocks similar to those intruding the New Bay Formation and Dunnage melange further suggests that the Burnt Bay chert is pre-Caradocian in age.

Relationship Between the Campbellton Sequence and the Caradocian Black Shale/Burnt Bay Chert Sequence

The Burnt Bay chert unit appears to pass upwards from a fine-grained sandy greywacke zone with interbedded argillaceous chert to green bioturbated cherts that characterize the Burnt Bay chert unit. This lower greywacke zone found in the southernmost portion of the Burnt Bay chert outcrop area is very similar to interbedded greywackes and shales found within both the Riding Island greywacke and the New Bay Formation. This implies that the Burnt Bay chert overlies both the New Bay Formation and the Riding Island greywacke and hence the Campbellton sequence. However, since outcrop between the Burnt Bay chert and Riding Island greywacke is sparse and the fact that greywackes from different units tend to look similar it could easily be argued that the Riding Island greywacke bears no relationship to the Burnt Bay chert and that the Caradocian black slate/Burnt Bay chert sequence is a lateral facies variation to the Luscombe Formation. This idea implies that the uppermost portion of the Luscombe Formation, composed of dark gray, highly argillaceous cherts (Member G) is temporally equivalent to the Caradocian black slate and that no graptolites have been found in this unit simply because it underwent hornfels-type metamorphism from the intrusion of the nearby Loon Bay batholith destroying any graptolites that may have existed here. The Burnt Bay chert is separated from the Luscombe Formation by the Campbellton fault. Displacement on the Campbellton fault cannot be large since the Riding Island greywacke is found on both sides of the fault. This suggests that there is 1 kilometer or less of some sort of oblique slip across the fault. If the Caradocian black shale/Burnt Bay chert sequence and the Luscombe Formation are correlative then the distance of separation between these units is much too small to account for the tremendous lithological differences through a lateral facies change. The Luscombe Formation gradually passes upward from a spessartite/rhodochrosite banded chert with the manganiferous minerals composing up to 30% of the total rock into a gradually more argillaceous and less manganiferous chert. The uppermost portion of the Luscombe Formation (Member G), consists of 50% shale and 50% chert. Conversely, the Burnt Bay chert appears to pass upwards from interbedded greywackes and shaly cherts into a very 'clean' chert (almost 100% microcrystalline quartz) with a local surficial manganese oxide stain interbedded with rare conglomerate and greywacke beds. This chert shows no gradual increase in shaly component towards the Caradocian black shale contact. There appears to be an abrupt transition from 'clean' shale free, chert into 100% shale across the Burnt Bay chert/Caradocian black shale contact; further suggesting that this sequence is not simply a facies variation of the Luscombe Formation.

Relationship Between the New Bay Formation and the Riding Island Greywacke

If the Campbellton sequence is pre-Llandeilian (pre-Lawrence Head volcanics) in age this implies that the Riding Island greywacke (uppermost unit of the Campbellton sequence) may be temporally correlative with the New Bay Formation that Helwig (1967, 1969) established to be of Llanvirnian/Arenigian in age.

Rocks very similar in appearance to the thin-bedded portion of the New Bay Formation are found in woodland outcrops of the Riding Island greywacke east of the Hales fault. A section very similar to the New Bay type greywackes found beneath the Burnt Bay chert in the southernmost portion of its outcrop area is found in the Riding Island greywacke along the Campbellton town dump access road. Thus, it appears that the Riding Island greywacke becomes more New Bay-like west of Campbellton towards Michael's Harbour. These relationships as well as the assumptions about the age of the Riding Island greywacke suggest that the Riding Island greywacke is identical to and a member of the New Bay Formation.

The Riding Island greywacke most likely represents a channelized sequence of quartz-rich turbidites that cut through and were deposited with the New Bay Formation.

Regional Stratigraphic Relationships

Stratigraphic sequences correlative with those found in the Lewisporte/Loon Bay area are seen in adjacent regions (Fig. 3-2). To the west along Dildo Run the Dunnage melange, containing rocks as old as Middle Cambrian, is overlain by the Cheneyville Formation and Dark Hole shale of Caradocian age (Kay, 1976). As discussed, these two units are lateral equivalents of the Burnt Bay chert/unnamed Caradocian black shale sequence found in my map area. The block of bedded green bioturbated chert identical with the Burnt Bay chert found incorporated into the Dunnage melange in the Dildo Run area implies that deposition of the Burnt Bay chert was partly diachronous with the formation of the Dunnage melange and farther supports Hibbard's (1976) idea that the Dunnage melange is the disrupted equivalent of nearby stratigraphic units.

To the west, in the New Bay-Fortune Harbour peninsula area, the type section of the New Bay Formation passes upwards into mafic pillow basalts of the Lawrence Head Formation (Helwig, 1967). Hibbard (1976) believes that most of the mafic pillow basalt blocks found within the Dunnage melange were derived from the Lawrence Head Formation. Above the Lawrence Head Formation is the Lawrence Harbour shale of Caradocian age. At the base of the Lawrence Harbour shale a thin (less than 1

Regional Stratigraphic Correlations

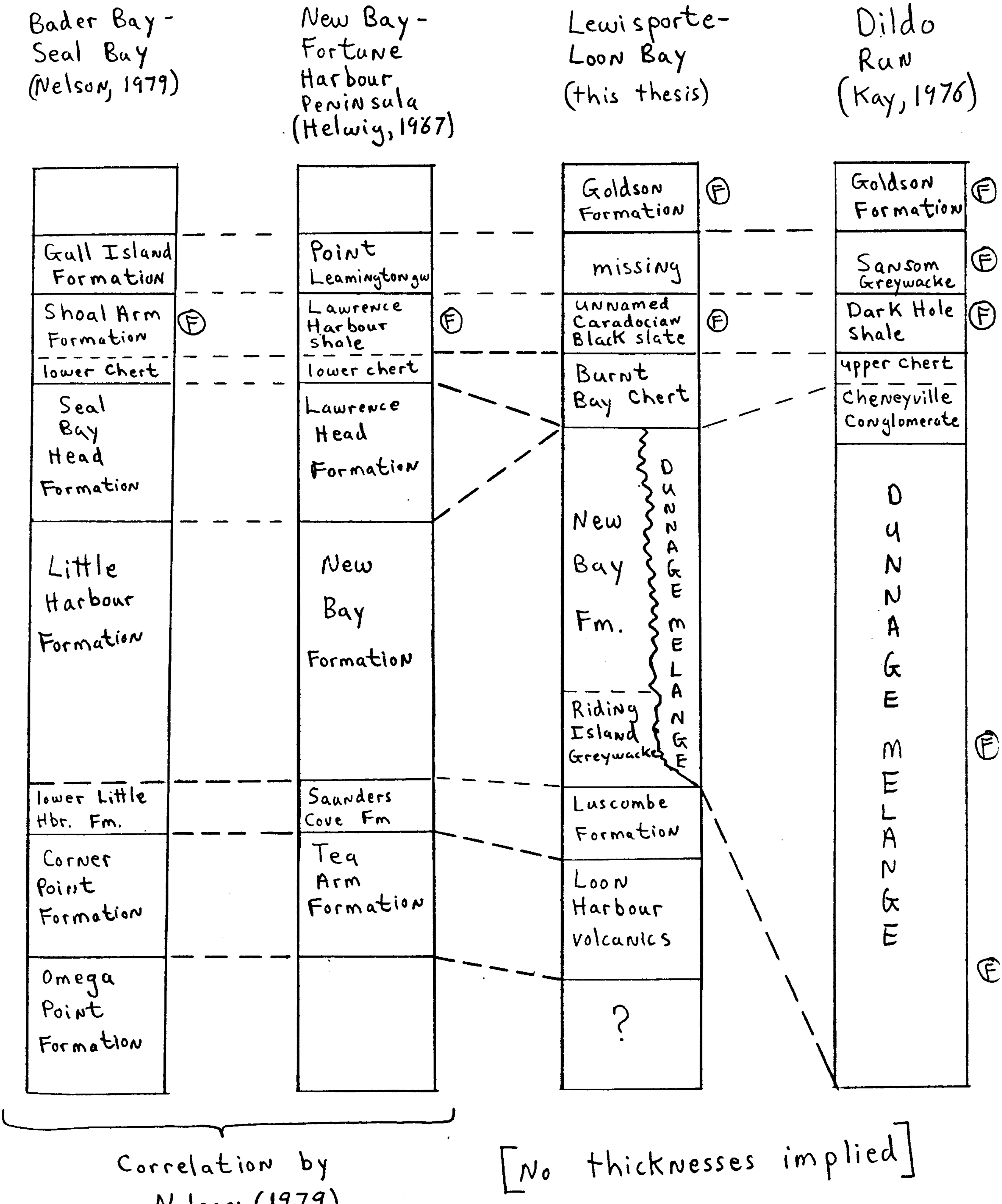


Figure 3-2

meter thick) layer of red and green chert is found (W.S.F. Kidd, pers. comm.). The Burnt Bay chert appears to be a distal lateral facies equivalent of this thin chert horizon since they both occur immediately beneath slates of Caradocian age. A similar chert is found at the base of the Caradocian age black shale of the Shoal Arm Formation (Nelson, 1979). Here a zone of red and green cherts with surficial manganese oxide stains passes upwards into a thin layer of green, bioturbated chert similar in appearance to the Burnt Bay chert. The red and green cherts at the base of both the Lawrence Harbour shale and Shoal Arm Formation are similar in appearance (except that the cherts at the base of the Lawrence Harbour shale do not have a surficial manganese oxide stain) and are most likely temporally correlative with the Burnt Bay chert.

An important difference between the Exploits Group and the stratigraphic sequence found in my map area is that the Burnt Bay chert passes downwards into the New Bay Formation (Riding Island greywacke) without any intervening mafic pillow basalt horizon as seen in the middle Exploits Group to the west.

If the Riding Island greywacke is a member of the New Bay Formation then the Luscombe Formation and Loon Harbour volcanics must lie beneath the New Bay Formation. Units that may be at least temporally correlative with the Luscombe Formation/Loon Harbour volcanics are found in the lower Exploits Group that lies beneath the New Bay Formation in the New Bay-Fortune Harbour peninsula area (see Helwig, 1967). The lower Exploits Group consists of the Saunders Cove Formation and underlying Tea Arm volcanics (Helwig, 1967). The Saunders Cove Formation is 515 m thick and consists dominantly of red and green, thin-bedded, ferraginous cherty argillite with minor tuff, conglomerate, greywacke and chert (Helwig, 1967). The thickness of the Luscombe Formation (370 m) is comparable to that of the Saunders Cove Formation. Also analogous is the argillaceous nature of the chert in both these units and the fact that both are interbedded with minor amounts of tuff, volcaniclastics, conglomerate and greywacke. Both units are also intruded by gabbro bodies. Of course, the distinct difference between these two formations is the lack of manganese-bearing minerals within the Saunders Cove Formation and its more argillaceous character. Nonetheless, the Saunders Cove Formation is better suited to be correlative with Luscombe Formation than any other unit within the Bay of Exploits. The lack of manganese-bearing minerals and the more argillaceous nature of the Saunders Cove Formation can easily be explained by a lateral facies variation since these units are separated by 25 km distance. Within the Luscombe Formation itself, a decrease in abundance of manganese-bearing minerals and a concomitant increase in the amount of argillite within the chert is seen from Loon Harbour westwards towards Campbellton. If this same trend were to continue westwards then one would expect that a lateral facies equivalent of the Luscombe Formation 25 kilometers to the west would contain no manganese-bearing minerals and to be composed of dominantly cherty argillite as is the Saunders Cove Formation. Correlation of the Luscombe and Saunders Cove Formation is further supported by the fact that both formations overlie mafic volcanic to volcaniclastic units. The Loon Harbour volcanics underlies the Luscombe Formation and is composed of mafic to intermediate volcaniclastic rocks with pillow basalt blocks and fragments occurring at the top of this unit. The Tea Arm Formation which underlies the Saunders Cove Formation is composed chiefly of mafic pillow lava, volcanic breccia, lava flows, agglomerate and minor tuff, chert and limestone (Helwig, 1967) with local horizons of silicic pillow lava and silicic volcanic breccia (Nelson, 1979). Although the Loon Harbour volcanics is composed of fragmental volcanic rocks, the lithologic makeup of both these units is similar implying that they are correlative.

Correlations between the lower and middle Exploits Group and the Wild Bight Group in the Seal Bay-Badger Bay area are given by Nelson (1979) and are illustrated on Figure 3-2.

CHAPTER IV

STRUCTURE

Introduction

The dominant structure of the Lewisporte/Loon Bay area consists of a series of moderate to steeply southeast-plunging inclined to reclined, close to tight, overturned macroscopic folds. Axial surface cleavage of these folds forms a regional penetrative cleavage that affects all rocks within the map area. These folds are interpreted to represent the first major phase of deformation that affected this area (D_1). Various local complications exist such as minor differences in style and orientations of the folds and associated lineations and cleavage. Some of the minor changes in fold orientations may be a result of a series of north-northeast trending sinistral faults that kink and offset cleavage. Important non-penetrative soft sediment structures pre-dating the major regional folding episode are found in several units.

The mapped area has been subdivided into four subareas (Fig. 4-1) on the basis of minor variations in the style and orientation of folds and related cleavages and lineations so that these structures could be described in detail in a systematic and orderly manner. Natural boundaries such as prominent faults, or more rarely stratigraphic contacts, are used to separate subareas that appear to have consistent internal structure that is somewhat different from adjacent subareas.

Subarea I

Subarea I includes the region west of the Hales Fault and north

Location map of Structural Subareas

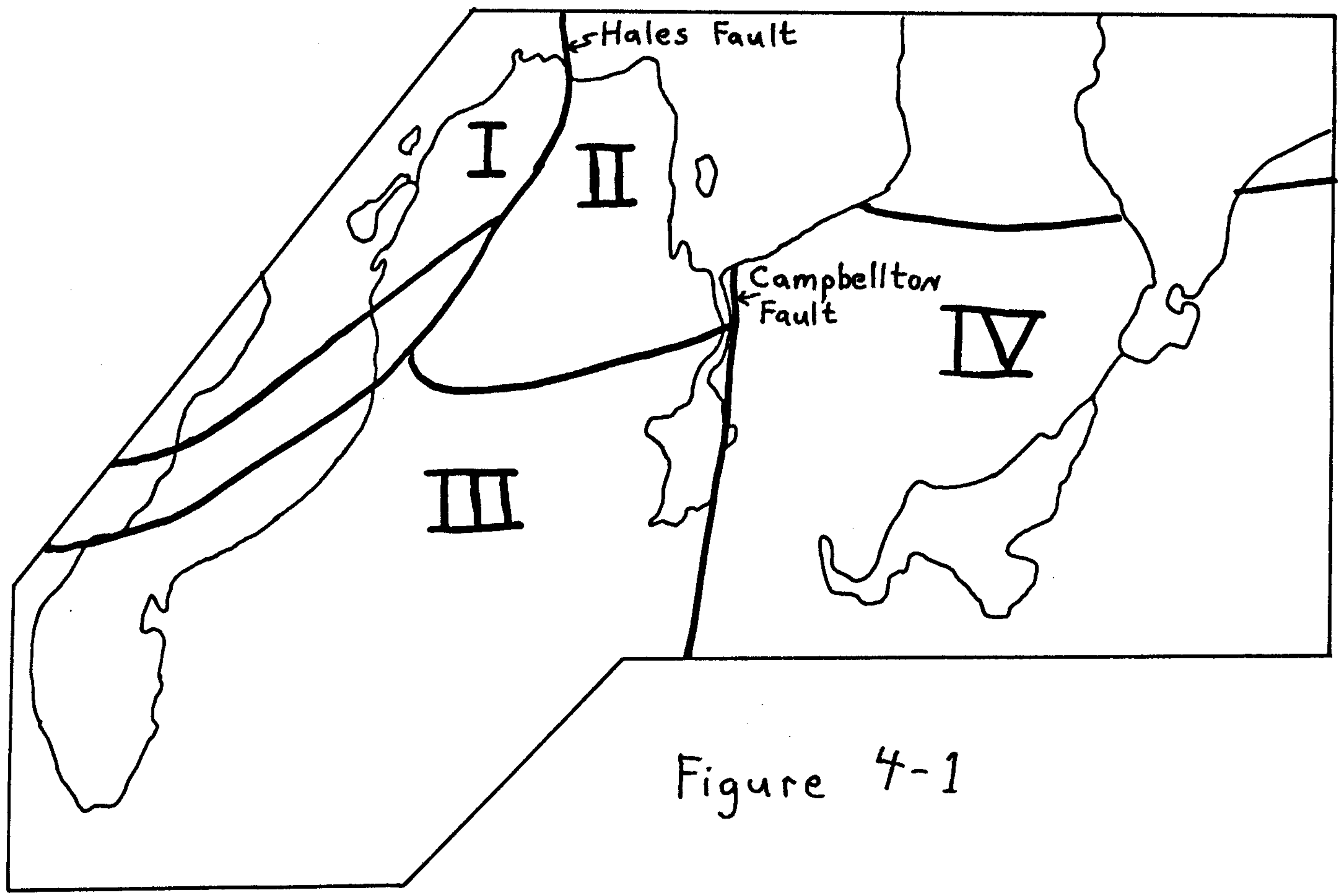


Figure 4-1

of the Botwood Group graben (see section on Faults). The New Bay Formation and Dunnage melange are found within subarea I.

A pervasive, well developed slaty cleavage (S_1) is found within the pelitic, thin-bedded portion of the New Bay Formation in subarea I (Fig. 4-3). In the very thick bedded conglomeratic horizons of the New Bay Formation a similar slaty cleavage (S_1) is found only within the numerous shaly rip-up clasts within these beds. This cleavage does not penetrate into the surrounding conglomerate. However, a second spaced cleavage (S_2) cleavage is found in this conglomerate that cuts through it and the slaty cleavage (S_1) within the rip-up clasts. The slaty cleavage is generally steeply inclined and found as an axial surface cleavage to numerous mesoscopic plunging inclined to reclined, close, similar-style folds. Hinge lines of these folds plunge towards the southwest. This slaty cleavage (S_1) intensifies westwards along the Michael's Harbour peninsula towards the Dunnage melange. Along the westernmost portion of Michael's Harbour this cleavage becomes a somewhat more lenticular slaty cleavage (S_1) that is parallel with the axial surfaces of similar style, tight folds that are best seen in thin, more competent carbonate interbeds within the thin-bedded argillaceous New Bay Formation. Local zones of more intense cleavage development exhibit isoclinal folds with boudinaged limbs and intrafolial fold hinges. Just to the south along the westernmost portion of the Emily Cove peninsula this same distinctive 'lenticular looking' slaty cleavage (S_1) is also found. Cleavage in this area does not reach the same intensity of development as found in the western portion of Michael's Harbour peninsula. Mesoscopic folds associated with cleavage in this area are symmetrical, open to close plunging inclined folds with wavelengths of less than 1/2 meter. Disrupted manganiferous buff-colored layers are found in the westernmost portion of Emily Cove peninsula as well as in the New Bay Formation along Rideout Point and the Dunnage melange along Mussel Bed Rocks. These layers have been chaotically boudinaged and close to isoclinally folded by polyclinal, similar type folds. These layers exhibit wispy boundaries with the surrounding argillaceous material. Hibbard (1976) described these features as 'toothpaste' structure that formed during the early 'soft rock' period of deformation that resulted from gravity-induced soft sediment slumping that essentially formed the Dunnage melange (Fig. 4-2, see also Plate 3 of Hibbard and Williams, 1979). These structures are found throughout the argillaceous matrix material of the southwest portion of the Dunnage melange and within the New Bay Formation along the New Bay Formation/Dunnage melange transition zone (Hibbard, 1976). This style of deformation is characteristic of the soft rock deformation seen throughout the entire Dunnage melange (Hibbard and Williams, 1979). Hibbard (1976), citing Wood (1974), claimed that the boudinage of the manganiferous buff-colored beds as well as of greywacke beds within the Dunnage melange had to be a soft rock process to achieve the needed ductility contrast between beds. Hibbard (1976) continually emphasized the 'soft rock' nature of the disruption of the middle Exploits Group to form the Dunnage melange and that this event predated development of the penetrative 'hard rock' cleavage that later modified the Dunnage. He further argued that several phases of 'soft rock' deformation were needed to produce the observed features (a good example of this is given by Hibbard and Williams, 1979, page 1016).

The soft sediment slump structures exhibited by the manganiferous

Figure 4-2. Chaotically folded buff-colored layers in New Bay Formation along Rideout Point in New Bay/Dunnage melange transition zone.

Figure 4-3: Well developed slaty cleavage (S_1) in New Bay Formation on Cat Island.

buff-colored layers are overprinted by the S_1 cleavage that must be a post-lithification structure because it forms augen around the manganiferous buff-colored layers and affects thin quartz veins that cut through this material. Similar evidence for the hard rock nature of this cleavage was given by Hibbard (1976).

This post-lithification S_1 slaty cleavage is also very well developed in the New Bay Formation on Cat Island where it forms axial surface cleavage to close, plunging mesoscopic folds that are locally tight to isoclinal.

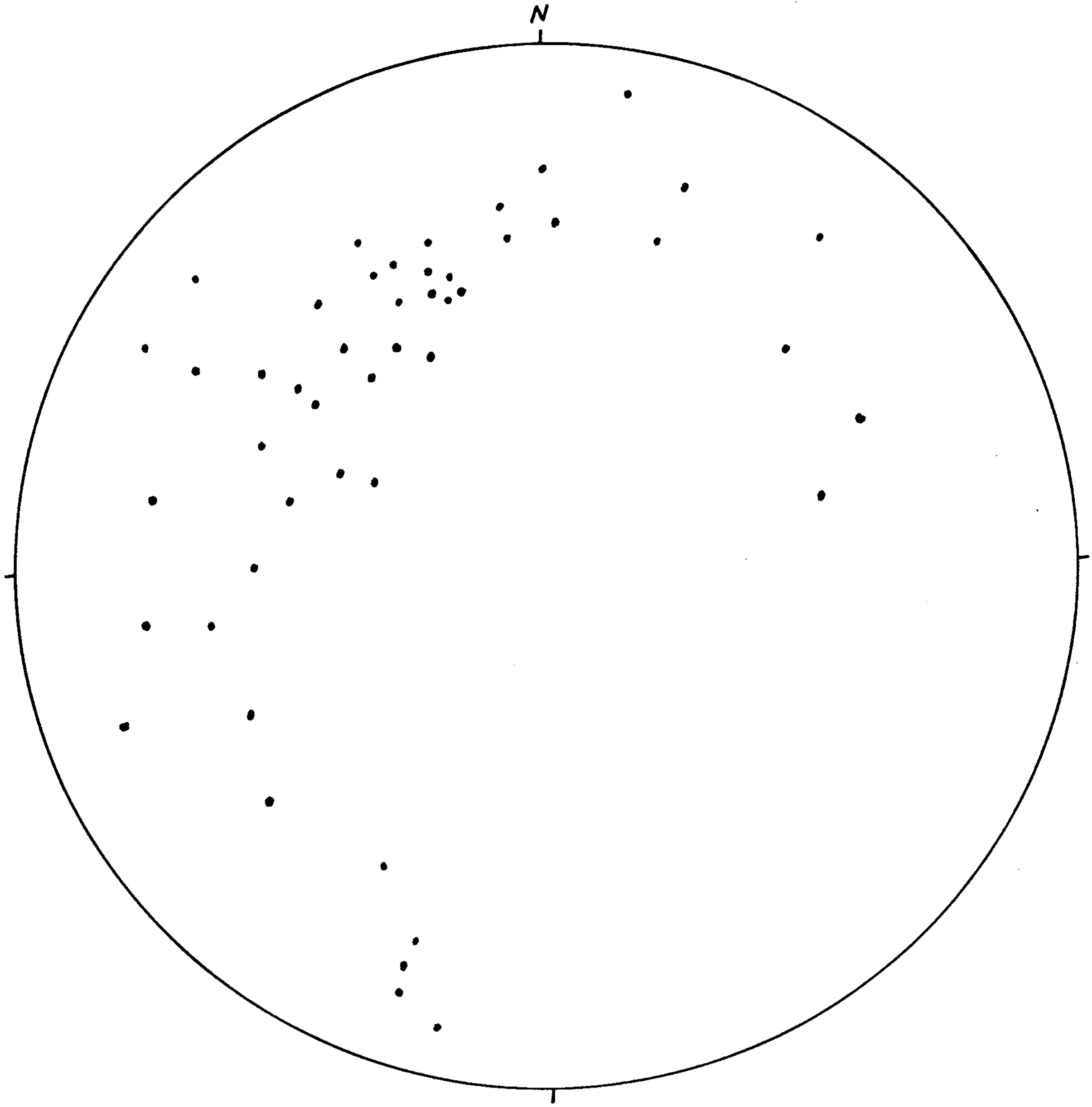
Along Mussel Bed Rocks within subarea I Dunnage melange is found. Structure is much more chaotic here than in the Dunnage melange along the western shore of Burnt Bay. The Dunnage melange in this area consists of strongly cleaved, locally very friable, pelitic matrix zones surrounding isolated blocks of pebbly greywacke up to 5 by 6 meters across. Cleavage in the pelitic matrix is a lenticular to phacoidal cleavage consisting of subparallel to somewhat chaotically oriented spaced cleavage surfaces. Manganiferous buff-colored layers are found throughout the rocks in this area exhibiting characteristic soft sediment disruption as previously discussed. These layers have been elongated and crenulated by later cleavage (S_1) development resulting in a distinctive, well-developed crinkle lineation to local elongation lineation (L_1) that is parallel with hinge lines of D_1 folds found throughout subarea I. This lineation is also parallel with the long axis of elongated pebbles in the New Bay Formation just to the east at Rideout Point. Cleavage offsets thin quartz veins on the eastern portion of Mussel Bed Rocks and forms augen around the isolated greywacke blocks implying that it is a 'hard rock' cleavage (see also Hibbard, 1976).

Since the Mussel Bed Rocks area is adjacent to the Loon Bay batholith most pelitic rocks contain cordierite porphyroblasts as a result of hornfels metamorphism. In thin section cleavage is deflected around the cordierite porphyroblasts. Inclusion trails found within the cordierite porphyroblasts are nearly perpendicular to the dominant external cleavage implying that porphyroblast growth predates the cleavage development. This further implies that cleavage in this area is a hard rock feature (Hibbard, 1976). Similar style, close to tight, mesoscopic folds are found in the Dunnage melange south of Mussel Bed Rocks towards Michael's Harbour. Hinge lines of these folds trend north-south and plunge moderately to the south. Since these folds are similar in their style and orientation to the pervasive plunging D_1 folds found throughout the map area they were assumed to represent F_1 folds.

Dunnage melange is also found just north of Cat Island along the western shores of Burnt Bay. Here, a single pervasive slaty cleavage (S_1) is found that augens intraformational pebbly mudstone melange clasts. Bedding (S_0) in this area is folded about open to close folds with the S_1 slaty cleavage as the axial surface cleavage implying that they are F_1 folds. Hinge lines of these folds trend north-south and plunge moderately to the south.

Figures 4-4 and 4-5 illustrate Lambert equal-area projections of poles to S_0, S_1 and F_1/L_1 lineations for subarea I. If it is assumed that the mesoscopic fold hinges and lineations are fabric elements of larger macroscopic folds then the pattern illustrated in figures 4-4 and 4-5 imply that a series of moderately southeast plunging inclined

Lower Hemisphere equal-area
Stereographic Projection

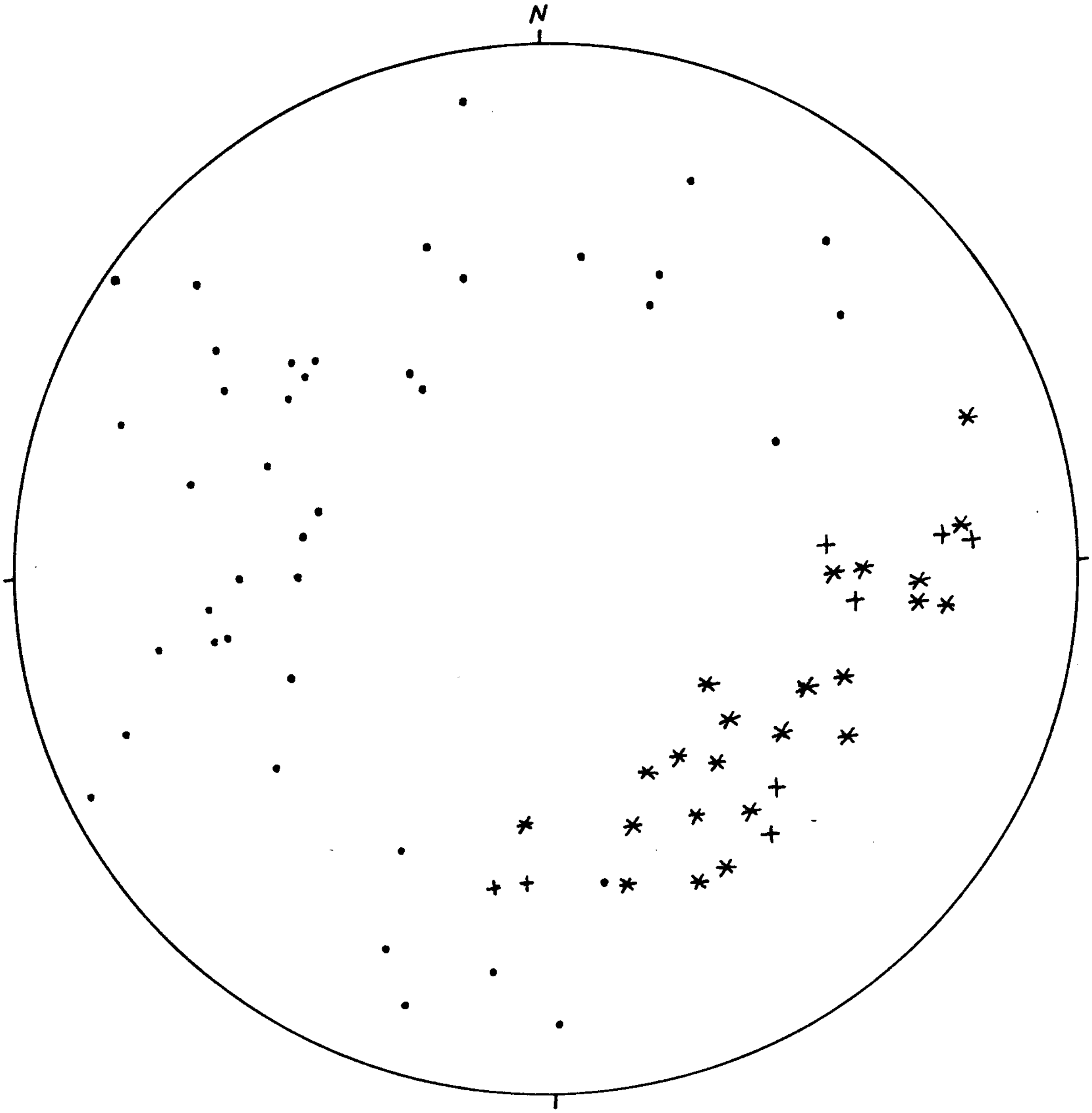


• poles to S_1

Subarea I

Figure 4-4

Lower Hemisphere equal-area Stereographic Projection



. poles to S_0

+ F_1 hinges

* L_1 lineations

Subarea I

Figure 4-5

to reclined macroscopic folds is found within subarea I. Overturned bedding found in this subarea suggests that the folds are overturned. The folds most likely vary from open to tight as the mesoscopic parasitic folds do (discussed further in Structural Evolution section).

Other small structures in this area include a set of conjugate shear fractures that consists of a very well-developed, vertical sinistral spaced fracture set (S_2). Quartz-filled sigmoidal extension gashes trending north-northeast (215) are associated with this set of fractures and a less well-developed dextral spaced fracture cleavage (S_2). Spacing on this fracture set is 1 cm or more with offsets of up to 6 meters to only a few centimeters. Rarely found conjugate kink bands trend southwesterly and are typically about 1 centimeter wide. Both of these small structures crosscut all previous structures. A well-defined conjugate kink band set was found offsetting the S_1 cleavage at Mussel Bed Rocks where a greenish-yellow gabbroic dike that strikes into one of these kink zones is offset by 6 meters.

Subarea II

Subarea II is located in the eastern portion of the Michaels Harbour/Campbellton peninsula and includes most of Campbellton proper (Fig. 4-1). Subarea II is bounded to the west and east by high angle faults. The northern boundary is the contact with the Loon Bay Batholith. The southern boundary of Subarea II was difficult to position due to a lack of outcrop and hence poor understanding of the structure but was arbitrarily placed along the Riding Island Greywacke/ Burnt Bay chert contact.

In general, bedding within Subarea II trends northeast-southwest and is moderate to steeply inclined (45-80 degrees) to the southeast varying from overturned to upright. In the southeastern portion of subarea II bedding in the Riding Island Greywacke and underlying Luscombe Formation is overturned, facing to the northwest and dipping to the southeast. Further north a facing reversal occurs as the Riding Island Greywacke is found to both face and dip to the southeast. This southeast dipping and facing relationship continues northward through the Riding Island Greywacke/Luscombe Formation contact just north of Campbellton. Within the Luscombe Formation north of Campbellton another facing reversal occurs as bedding again becomes overturned and Member C (highly manganiferous zone) of the Luscombe Formation is repeated. This overturned zone of the Luscombe Formation is then juxtaposed against New Bay Formation whose bedding strikes northeast/southwest similar to the Luscombe Formation, however, bedding is not overturned but right side up.

Mesoscopic folds are well displayed within the Luscombe Formation north of Campbellton. Here close, concentric folds and local parallel style folds with wavelengths of 5 cm to 4 meters are found. A poorly developed axial surface cleavage is usually associated with these mesoscopic folds. This cleavage is a slaty cleavage (S_1) in the argillaceous portions of the rock passing into a fine- to medium-spaced (spaced less than ½ cm apart) cleavage in more cherty zones. In general, the hinge lines of these folds plunge moderately to steeply (55-80 degrees) to the southeast (310-350 degrees).

Parallel with the hinge lines of these mesoscopic folds is a lineation seen in the cherty members of the Luscombe Formation (Fig. 4-8).

This lineation is a well-developed ribbing that looks like very small scale folds (on the order of mm in wavelength) that are laterally continuous. They are somewhat similar to crinkle lineations in shaly rocks except that they do not crenulate any earlier cleavage and, of course, they are found in very cherty rocks. No mineral or structural lineation was seen in thin sections of these rocks that were cut at several orientations to the lineation. Only pervasively aligned mica flakes defining the S_1 cleavage fabric was seen. This lineation could represent the crenulation of bedding by the S_1 cleavage or it could be an elongation lineation related to the folding. These lineations (L_1) are northwest trending (between 310 and 360) and plunge steeply to the southeast (55-80^o). When associated with mesoscopic folds (F_1) this lineation is parallel with the hinges of the folds.

Within the Riding Island Greywacke a locally well-developed cleavage (S_1) is recognized. This cleavage is a slaty cleavage in the argillaceous portions of the Riding Island greywacke refracting into the greywacke portions of this unit as a spaced cleavage spaced 1 cm or more apart. When seen the orientation of this cleavage is similar to the attitude of the axial surfaces of the mesoscopic folds found within the Luscombe Formation. Therefore, on the basis of similarity of orientation and the fact that this cleavage is the first generation both in the Luscombe Formation and Riding Island Greywacke it is assumed that these cleavages are both of the same generation and are axial surface to macroscopic folds.

In the northern portion of subarea II along Rideout Point conglomerate beds of the New Bay Formation are found. Pebbles in these conglomerates are strongly elongated. The orientation of the maximum elongation axis of these pebbles is parallel with the observed mesoscopic fold hinges, lineations and bedding/cleavage intersection lineations (Fig. 4-6). This is a commonly relationship for elongated pebbles and reflects elongation parallel with the fold hinges. If the pebbles are assumed to have been originally spherical and if they had the same mechanical properties of the surrounding matrix their shape will equal the strain ellipsoid in a homogeneous strain (Hobbs and others, 1976).

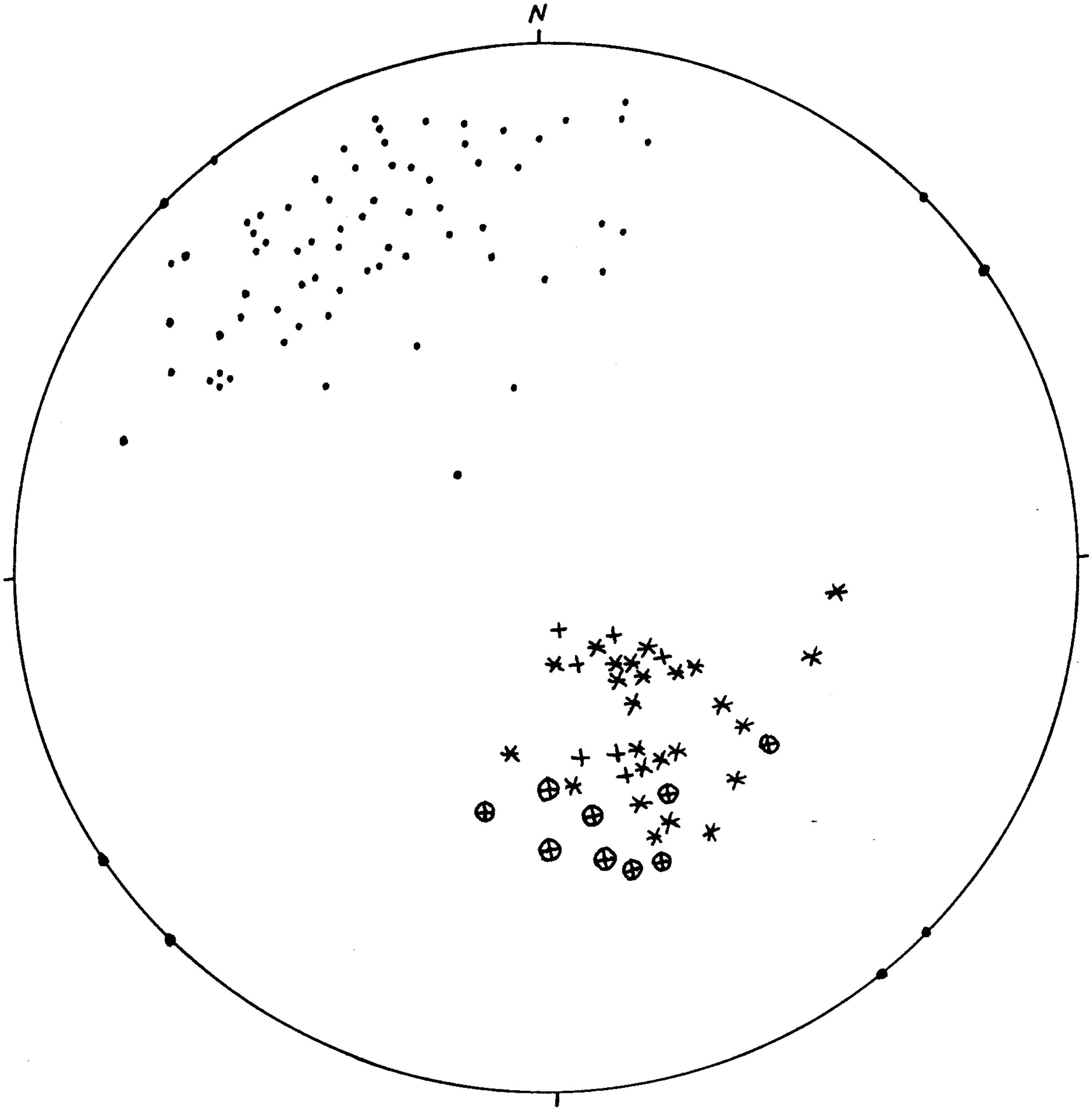
Three-dimensional ellipsoidal shapes of the deformed pebbles were measured from a cut slab of conglomerate collected along Rideout Point. Measurements taken on a few pebbles in this slab gave X:Y:Z axial ratios that range from 20:4:1 to 4:2:1 giving K values of 1.3 to 3 or greater than 1 (elongation field). The variation in the amount of strain measured for the pebbles is a function of lithologic type. More competent "granitic" pebbles are less strained whereas very ductile argillaceous pebbles are strongly strained and elongated. The most important drawbacks to this analysis is that the original shape of the pebbles are not likely to have been exactly spherical and that the mechanical properties of the pebbles will be vastly different from the surrounding matrix. Also, the each individual pebble may have experienced its own local strain history that will differ from that of the controlling folds (Hobbs and others, 1976). Pebbles may become elongated parallel with hinges of folds by rotating into this orientation (Jeffery, 1923). Ramsay (1967) claims that the initial shapes of the pebbles strongly influences its strain history. Originally elongate pebbles may orient themselves parallel with the hinge of the fold even if the maximum elongation direction is perpendicular to the fold (Ramsay, 1967; see also Hobbs and others, 1976, pages 284-286). Therefore, the best that can be said about the folds in subarea II is that some component of elongation (not necessarily maximum elongation) is found parallel with the hinges of the folds.

In the argillaceous portions of the New Bay Formation along Rideout Point folded buff-colored, thin spessartite-bearing layers are found. These layers appear to have been boudinaged in a ductile fashion and are folded about close to isoclinal, similar-type, polyclinal folds suggesting that much of this deformation is a result soft-sediment slumping (discussed in subarea I in more detail).

Mesoscopic structures mapped within subarea II, such as fold hinges, bedding/cleavage intersection lineations and elongation lineations all have similar attitudes plunging moderately to steeply towards the southeast (Figs. 4-6 and 4-7). If these mesoscopic structures are fabric elements of larger macroscopic folds then the preferred orientation of these structures as illustrated in Figures 4-6 and 4-7 as well as the facing reversals observed along Campbellton imply that this area is characterized by a series of north-northeast trending moderately to steeply inclined to reclined folds. The pattern illustrated in Figure 4-6 also suggests that this deformation was somewhat homogeneous..

Crosscutting and locally obliterating the S_1 cleavage is a spaced S_2 cleavage (spaced from 1 cm to 10 cm apart) striking approximately 330^o and dipping approximately 80^o to the southwest. Apparent horizontal dextral offsets of up to 20 cm are observed along this spaced (S_2) that probably represents the dextral strike-slip half of a post-folding minor conjugate shear system found throughout the entire mapped area.

Lower Hemisphere equal-area
Stereographic Projection



. poles to bedding (S_0)

+ F_1 hinges

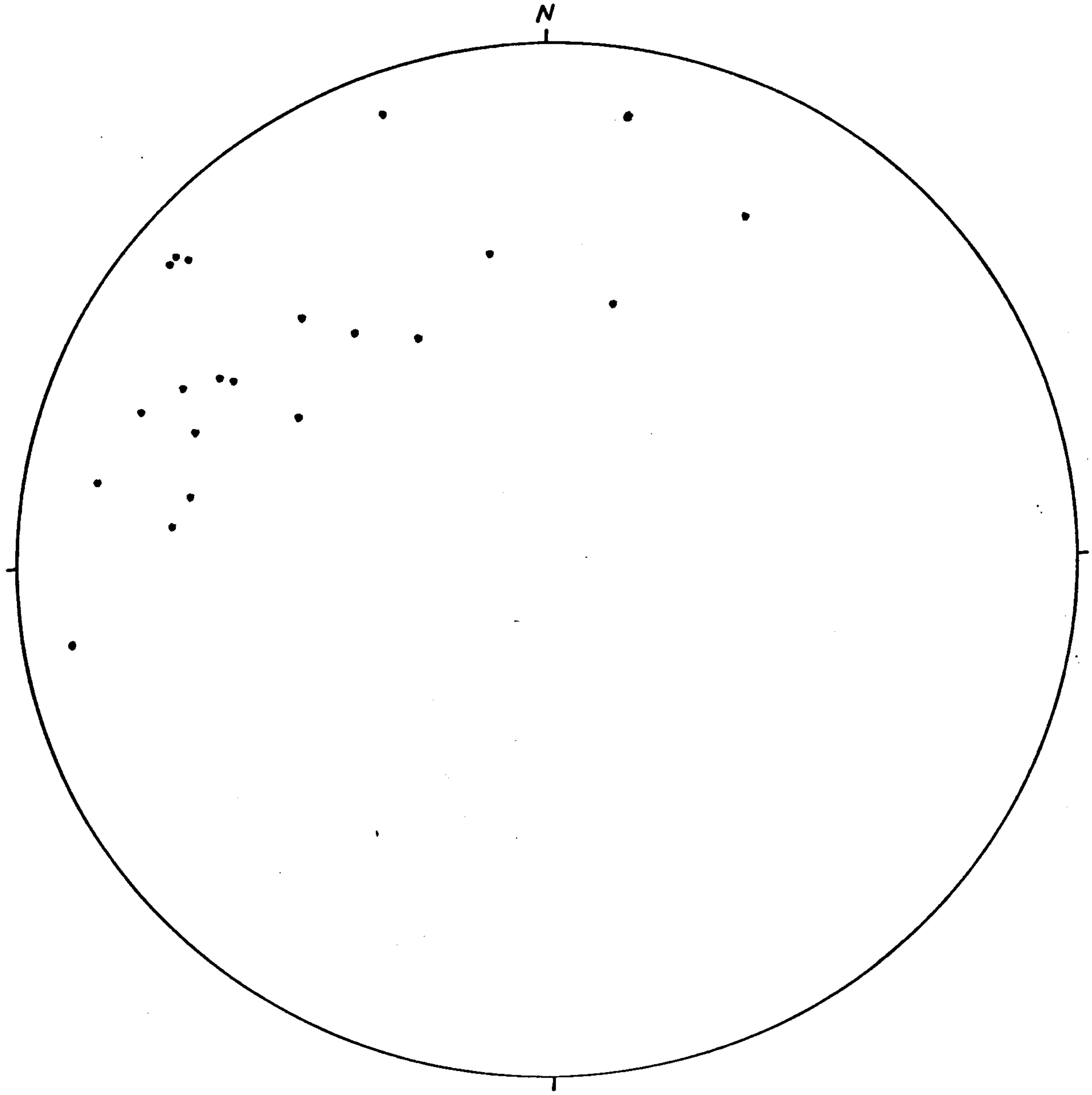
* L_1 lineations

⊕ Pebble Elongation lineations (L_1)

Subarea II

Figure 4-6

Lower Hemisphere equal-area
Stereographic Projection



• Poles to S_1 Slaty Cleavage

Subarea II

Figure 4-7

Figure 4-8. Lineation in manganiferous chert of the Luscombe Formation just north of Campbellton.

Subarea III

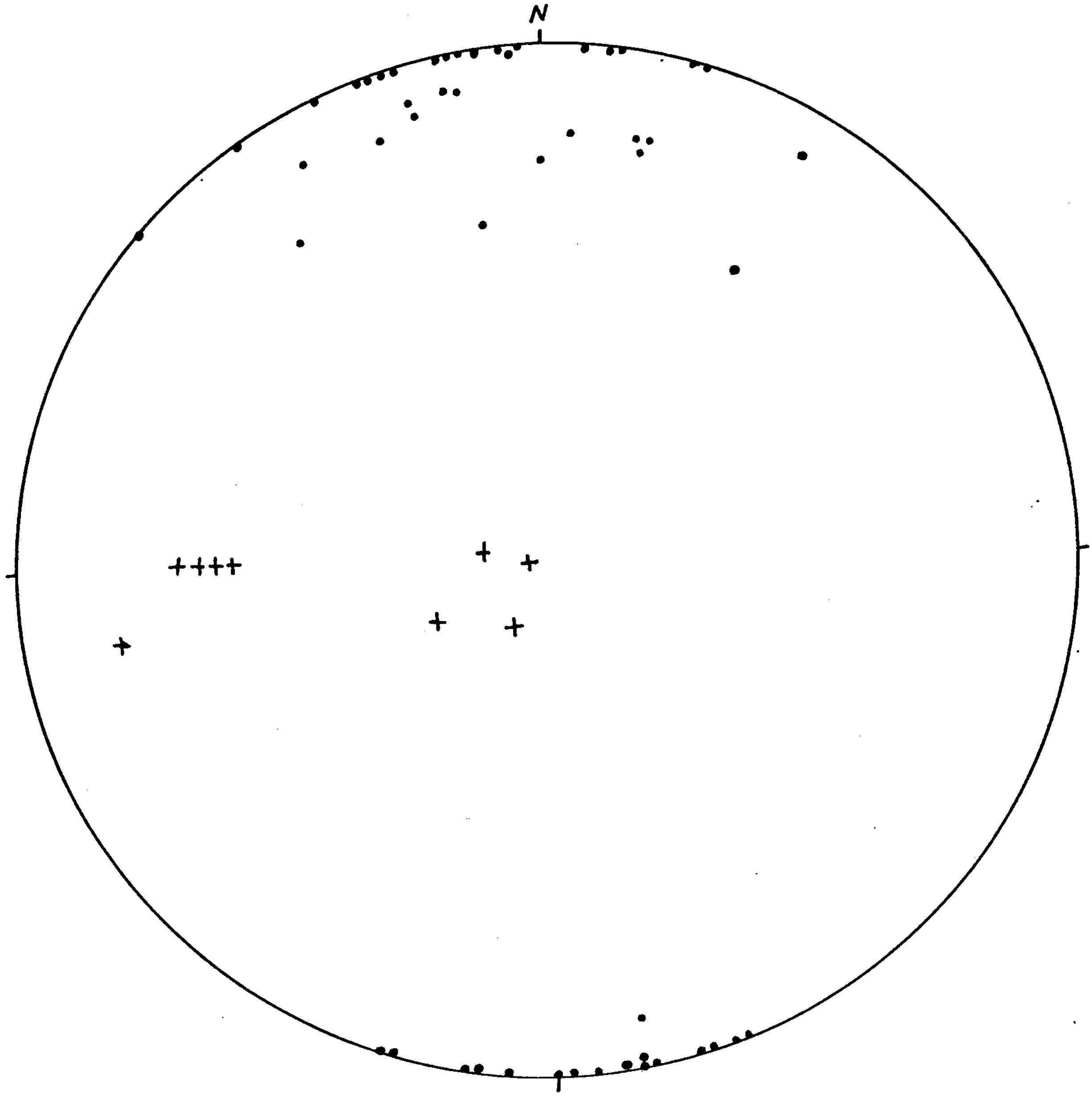
Subarea III includes the region where outcrops of the Burnt Bay chert are found. This subarea is bounded to the south, west, east and northwest by faults that down-drop younger Silurian and uppermost Ordovician rocks against the Burnt Bay chert. The northeast boundary of this subarea was arbitrarily placed along the contact of the Burnt Bay chert and the Riding Island Greywacke.

Found within subarea III are numerous mesoscopic close, upright, concentric folds with rounded hinges that trend east-west and plunge from 20^{o} west to vertical. A well developed, finely-spaced axial surface cleavage (S_1) (spaced < 1 cm apart) is associated with these folds within the chert beds. This cleavage generally strikes east-west and is vertically dipping. A similar finely-spaced cleavage (S_1) with the same orientation found in other chert outcrops where no mesoscopic folds are recognized and is assumed to be an axial surface fabric associated with macroscopic folds of the same style as the mesoscopic folds.

Sparsity of outcrop within Subarea III does not permit a clear understanding of the structure. However, stereographic plots of poles to both bedding and cleavage, mesoscopic fold hinges, and bedding/cleavage intersection lineations suggest the presence of east-west striking plunging upright to vertical close folds (Figs. 4-9 and 4-10).

Cross-cutting the S_1 spaced cleavage is a conjugate set of more coarsely spaced vertical fractures (S_2). These cleavages are generally spaced between 1 to 20 cm apart with one set striking approximately 195^{o} and the other striking approximately 345^{o}. Sinistral strike-slip offsets of up to 1 meter are found along the set that trends 195^{o}

Lower Hemisphere equal-area
Stereographic Projection



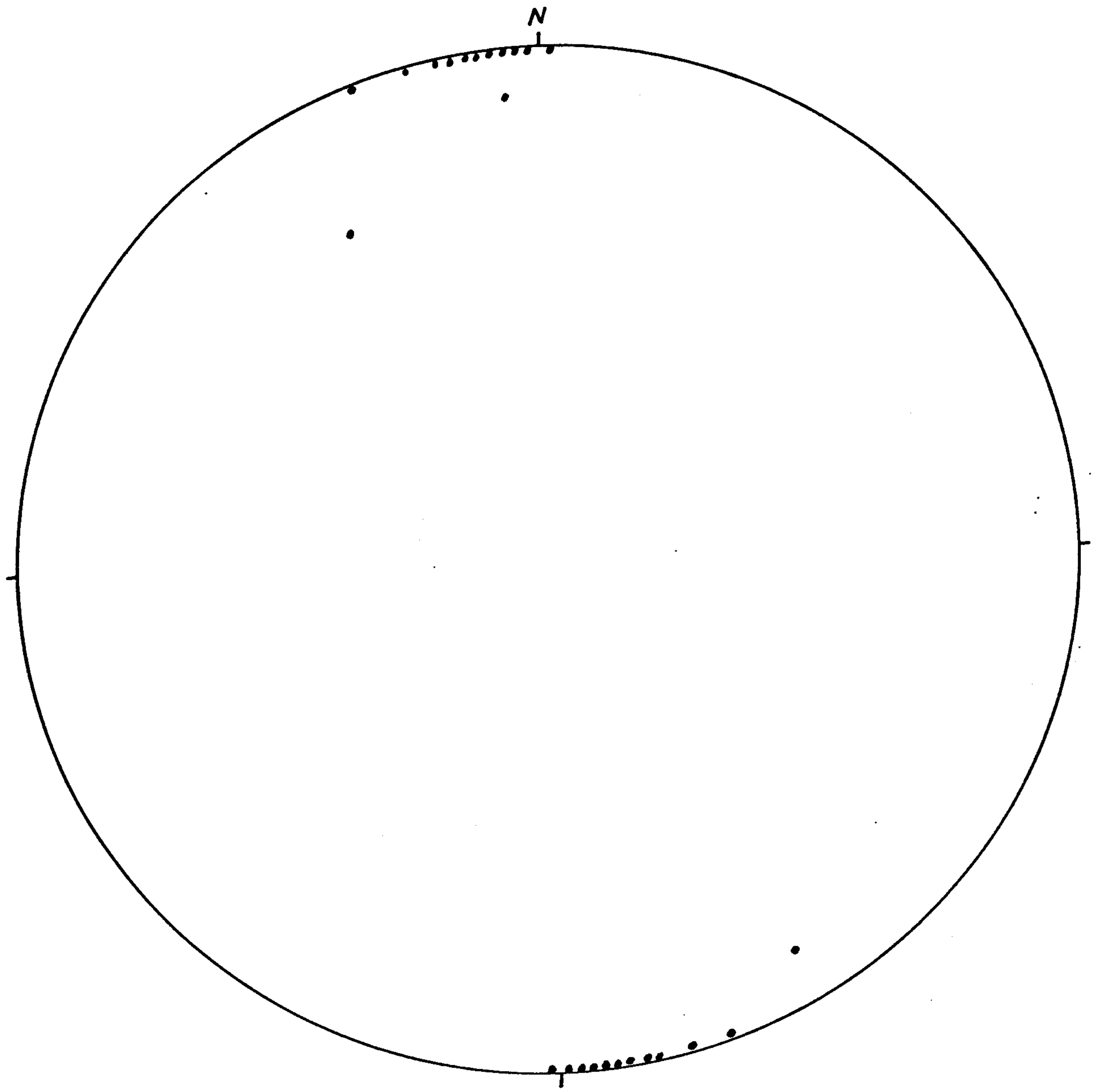
• poles to S_0

+ F_1 hinges and L_1 lineations

Subarea III

Figure 4-9

Lower Hemisphere equal-area
Stereographic Projection



Subarea III

• poles to S_1 (slaty cleavage)

Figure 4-10

implying that the S_2 cleavage represents a conjugate shear fracture set as also found in other portions of the map area.

Subarea IV

Subarea IV includes the region between Loon Harbour and Campbellton. Rock units found within subarea IV are the Loon Harbour Formation, Luscombe Formation and Riding Island greywacke. Subarea IV is bounded to the south by an east-west trending fault that downdrops the younger Goldson Formation against the Loon Harbour Formation. The intrusive contact of the Loon Bay Batholith forms the northern boundary of subarea IV. The western boundary of subarea IV is formed by a high angle strike-slip fault located in Campbellton along Indian Arm Brook. The eastern boundary is simply the edge of the mapped area.

As discussed in the section on Faults (see below), this area contains numerous north-northeast trending sinistral strike-slip faults that cross-cut and disrupt any macroscopic folds. Amount of outcrop and structural data are fairly abundant in the western and eastern portions of subarea IV. Unfortunately, outcrop is rather sparse in the central portion of subarea IV making structural interpretations difficult.

Bedding (S_0) in this area generally strikes east-west and is steeply dipping to vertical with stratigraphic units usually facing north. Bedding is overturned near Campbellton.

Numerous mesoscopic plunging inclined close to tight folds are found within the Luscombe Formation. The hinge lines of these folds generally trend northwest-southeast (approximately 305) and are moderately plunging to the southeast (approximately 50 SE). Axial surface cleavage of these folds varies from a weak to a well developed, fine to coarsely spaced cleavage in cherty rocks to a variably developed slaty cleavage in more argillaceous rocks. Stereographic plots of the hinges of these folds (Fig. 4-11) show that they all plunge to the southeast.

A locally well-developed, slaty cleavage is found both in the volcaniclastic layers and the more argillaceous portions of the Luscombe Formation and the Riding Island greywacke. This cleavage refracts through the cherty portions of the Luscombe Formation as a fine- to coarsely-spaced cleavage (spaced < 1 cm apart to 5 cm apart). Within the Loon Harbour volcanics a weakly developed, slaty cleavage is rarely seen. Due to the numerous later spaced cleavages slaty cleavage (S_1) in subarea IV generally strikes from east-west to northwest-southeast-dipping, steeply to vertically dipping (Fig. 4-12). Since a slaty cleavage is found as an axial surface cleavage to mesoscopic folds and since all of the slaty cleavage is roughly of the attitude it is assumed that all of the slaty cleavage within subarea IV is axial surface to the controlling macroscopic folds.

Coaxial with the F_1 fold hinges a well developed elongation lineation is found in the cherty portions of the Luscombe Formation and in the conglomerates of the Loon Harbour Formation.

Thin sections of lineated chert samples oriented at various angles to the lineation showed no elongated minerals or crenulation folds that could cause this lineation. This could be due to the fact that these rocks have been highly hornfelsed which could obliterate the microscopic evidence of this lineation.

Within the pebbly portions of the Loon Harbour Formation an elongation lineation is found whose attitude is parallel with the observed mesoscopic fold hinges. This lineation is best seen along the southern portion of Luscombe Point in Loon Harbour where basaltic pebbles are strongly elongated with axial ratios of approximately 5:1 (X:Y) as measured on several samples in the field.

If it is assumed that the mesoscopic structures are fabric elements of larger macroscopic folds then stereographic plots of F_1 hinge lines, L_1 lineations and poles to bedding (S_0) and cleavage (S_1) suggest that the macroscopic folds responsible for the structure in subarea IV consist of a series of overturned, plunging, inclined to reclined close folds. One of these folds may be found in the extreme northeast corner of subarea IV where repeated stratigraphy of the Luscombe Formation (using the distinctive member C as a key horizon) is found. If this is the case, the wavelength of the fold is between ½ to 1 ½ km. Along the western shore of Loon Harbour north of Brenson Point bedding in the Luscombe Formation strikes north-south deviating from the overall east-west trend of the strata within subarea IV. This may be due to either disruption of bedding by the later north-northeast trending strike-slip faults or it may represent the hinge of one of the macroscopic F_1 folds.

Cross-cutting and sometimes obscuring the earlier S_1 slaty cleavage are two sets of vertical, coarsely-spaced shear fractures (S_2). One of these fractures trends NNE (200-190) along which apparent sinistral offsets of up to 4 cm are seen in places. The other spaced fracture is less well developed and trends NNW (320-350^o). Since the sinistral spaced fracture is parallel with the NNE trending

Lower Hemisphere equal-area
Stereographic Projection

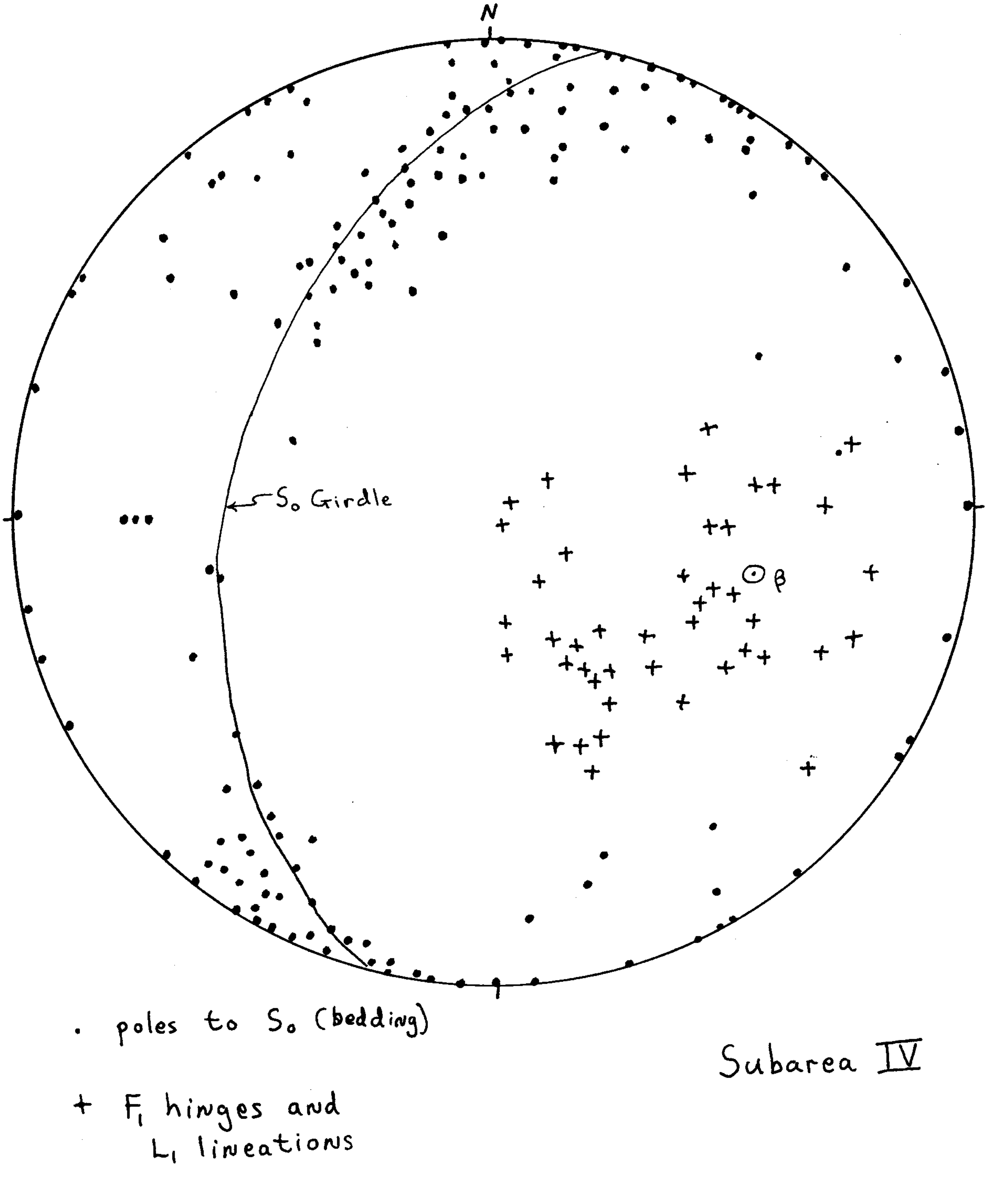
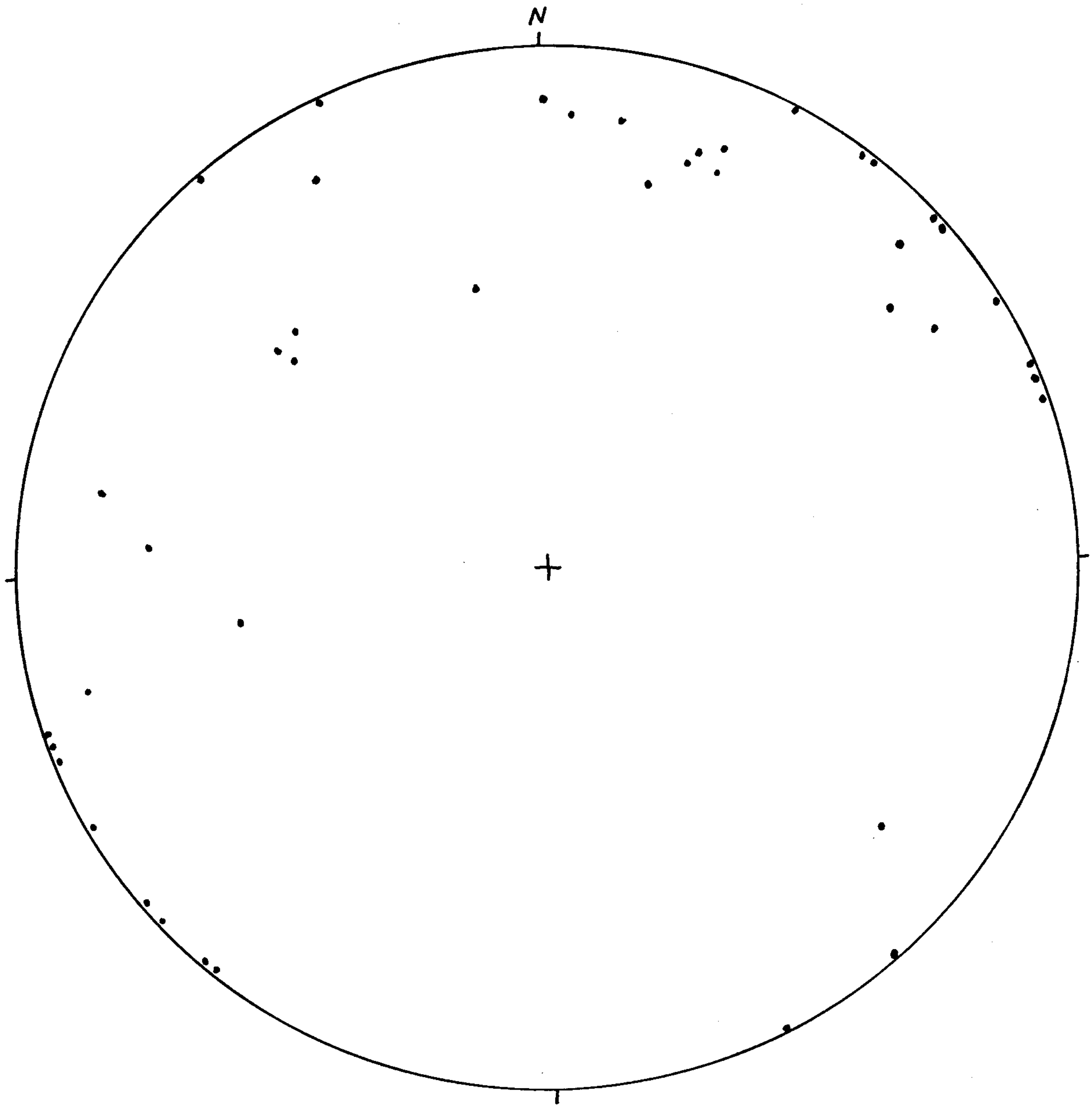


Figure 4-11

Lower Hemisphere equal-area Stereographic Projection



. poles to S_1 (slaty cleavage)

Subarea IV

Figure 4-12

set forming a conjugate shear couple with the NNE sinistral shear cleavage.

Faults

Numerous en-echelon north-northeast striking faults are found between Loon Harbour and Campbellton. Obvious sinistral offsets of generally vertically-dipping stratigraphic units suggest that these faults formed by mainly sinistral strike-slip with a lesser component of dip-slip. Along the western shore of Loon Harbour breccia zones several meters wide are found trending parallel with these faults. Stratigraphic separation across these faults is up to ½ kilometer. Associated with these north-northeast trending, dominantly strike-slip faults are a series of small, northwest to east-west striking, splay faults found along the western portion of Loon Harbour. These faults are identified by the trends of the breccia zones and obvious stratigraphic offsets.

In the Campbellton area a similar north-northeast striking, dominantly sinistral, strike-slip fault is found trending parallel with the northernmost portion of Indian Arm Brook. Just west of and parallel with this fault is another fault termed here as the Campbellton Fault. The amount displacement across this fault is not certain but cannot be great because the Riding Island greywacke is found on both sides of the fault.

These faults are parallel with the Reach, Burnt Arm and Virgin Village Faults found on New World Island and in Dildo Run (see Kay, 1976). Kay (1976) mapped sinistral displacements of up to 1 km on the Burnt Arm and Virgin Village faults. Since these faults have the same trend and sense of slip as those found within my map area, they are assumed to be related. This further implies that the Reach fault is a large scale sinistral strike-slip fault.

Striking north-south through the middle of the coastline of the Michael's Harbour/Campbellton peninsula is the Hales Fault identified by Kay (1975). The Hales Fault appears to continue southward to form the southern fault boundary of a northeast-southwest graben that down-drops the Silurian Botwood Group against older units. In the Michael's Harbour/Campbellton peninsula, rocks along the Hales Fault contain a single, pervasive, north-south trending, finely-spaced, lenticular cleavage that is parallel with the trend of the Hales Fault. Therefore, this fault may represent a tectonic slide that cut through the axial surface of a macroscopic D_1 fold. Displacement on the Hales Fault is uncertain but must be significant since the Luscombe Formation and Riding Island Greywacke are found along the eastern side of this fault but not the western side of it.

Kay (1975) speculated that an east-west trending, low angle thrust fault that was subsequently rotated to a subvertical attitude separated the New Bay Formation found along Rideout Point from the Luscombe Formation just to the south. Evidence for this fault is seen in an east-west trending lineament that can be traced on air photos from the shore westwards into the woodland area where the lineament gradually dies out. Across this fault beds in the New Bay and Luscombe Formations face towards each other, further implying that these units are separated by a fault.

In the western portion of the mapped area a northeast-trending graben downdropping the Silurian Botwood Group against older units is found. This graben trends east to northeastward from the Bay of Exploits (see Deans unpublished maps) across Burnt Bay and ends between Michael's Harbour and Campbellton. Tracing the extent of this graben was not difficult due to the distinctive lithologic nature of the Botwood Group. The northernmost fault of this graben is exposed in a roadcut along Route 340 just south of Michael's Harbour where several friable, phacoidally cleaved and brecciated zones are found within the Lawrenceton Formation of the Botwood Group. These zones strike northeastwards (between 220-210°) and dip 55° to the southeast, suggesting normal displacement on these faults.

The northernmost fault bounding this graben trends southwestward into Burnt Bay through Seal Rock, identified by Hibbard (1976) as quartz-rich brecciated fault rock. This fault is also exposed within Freak Island where brecciated outcrops of the Lawrenceton Formation are exposed that contain angular breccia fragments up to 15 cm across. This graben appears to be terminated between Michael's Harbour and Campbellton where outcrops of Lawrenceton Formation gradually die out giving way to outcrops of the New Bay Formation. It is assumed that the bounding faults of the Botwood Group graben pass into the Hales Fault since the inferred position of the Hales Fault appear to pass through the region where outcrops of the Botwood Group end. Outcrops of both the New Bay and Lawrenceton Formations are characterized by multiple cleavages and associated crinkle lineations proximal to the Hales Fault and bounding faults of the Botwood Group graben.

In the eastern portion of the mapped area, pebbles in the Goldson Formation near the contact with the Loon Harbour Formation are strongly elongated and cut by shear fractures to form a well-developed pebble elongation lineation (K > 1; Flinn, 1962). As measured on several samples in the field two dimensional axial ratios (X:Y) of the elongated pebbles vary from 10:1 to 3:1 depending on the lithology of the pebbles. Argillaceous pebbles are more highly elongated than quartzite and granitic pebbles. Also, some pebbles have been sliced along shear fractures that are at low angles to the principal elongation direction of the pebbles. Offsets along these shear fractures is less than ½ cm. Further south, into the Goldson Formation this pebble elongation fabric is not found. Since pebbles in the Goldson Formation are deformed only near the contact with the Loon Harbour Formation, it is inferred that this contact is an east-west striking fault contact that downdrops the Goldson Formation against the older Loon Harbour Formation.

Since the Goldson Formation appears to be in fault contact with the Loon Harbour Formation in the eastern portion of the map area, it was inferred that this formation is also in fault contact with the Burnt Bay chert in the western portion of the mapped area.

Age and cross-cutting relationships between the east-west striking fault that downdrops the Goldson Formation against older units and the north-northeast striking sinistral strike slip faults is not clear simply because of a lack of outcrop. However, it was inferred from the limited outcrop patterns that the east-west trending fault that downdrops the Goldson Formation against older units is older than the north-northeast striking sinistral strike-slip faults and is offset by these strike-slip faults. Also, the fact that the Goldson Formation strikes into the Burnt Bay chert in the western portion of the mapped area suggests that these units are in fault contact.

Structural Evolution

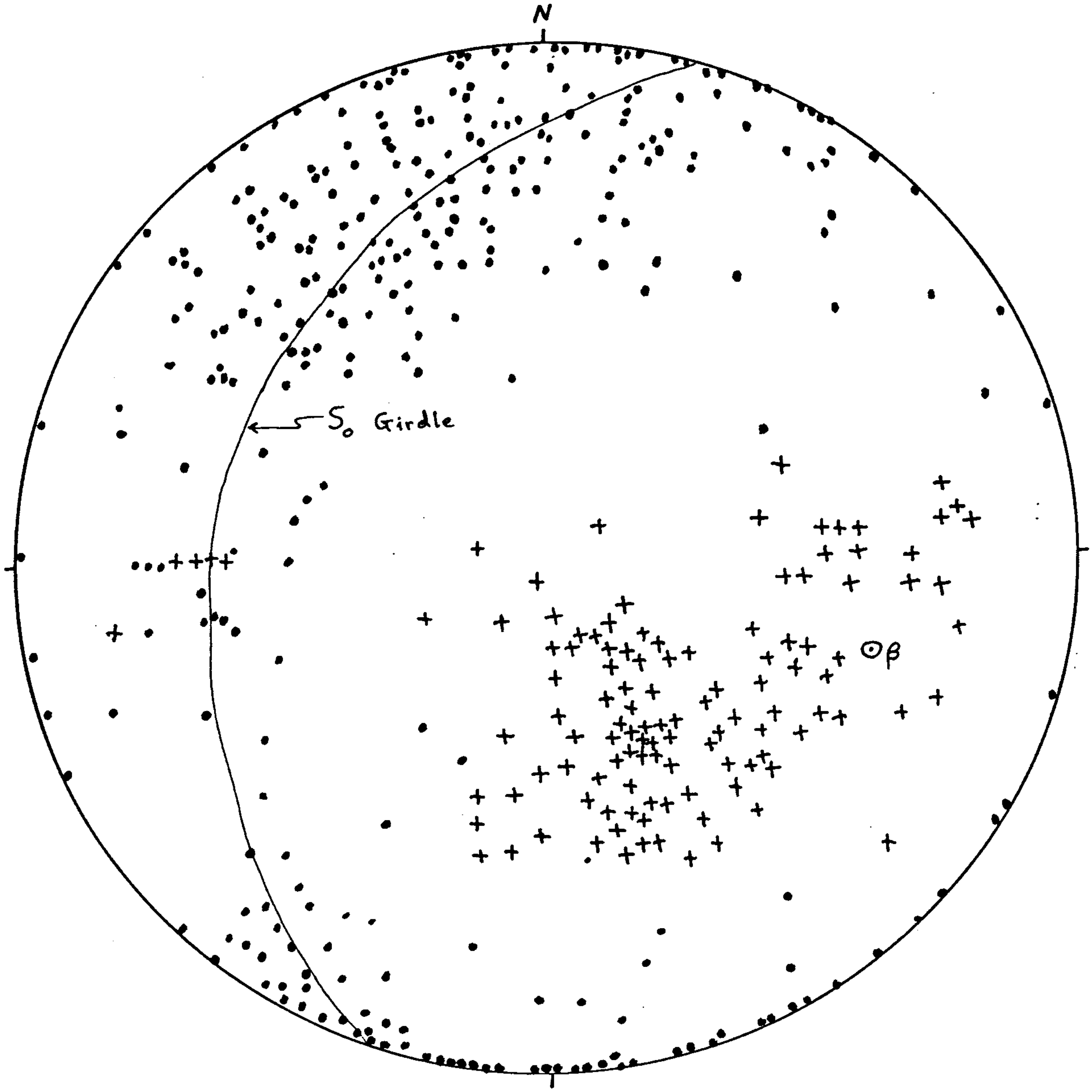
A north-northeast and north-northwest sinistral/dextral shear fracture set recognized throughout the map area forms a conjugate spaced fracture (S_2) set that crosscuts the S_1 slaty cleavage associated with the F1 folds. The sinistral set is more highly developed than the dextral set. Since the trend of the north-northeast trending en echelon sinistral strike-slip faults are identical with the trend of the sinistral shear fracture (S_2) it is assumed that they are genetically related. Therefore, the north-northeast trending faults must be post-F_1 structures. These faults appear to be representatives of the fault set containing the Reach, Burnt Arm and Virgin Arm faults. Kay (1976) claimed that the Loon Bay batholith was deformed and truncated by the Reach Fault. If this is the case, than these faults are either post-Silurian or post-Devonian in age (see discussion on age of Loon Bay batholith in Age of Campbellton sequence section). Horne (1968) recognized a similar north-northeast trending set of sinistral strike-slip faults on New World Island (Virgin Village fault and other smaller faults to the west) that also crosscut the regional folds in that area.

The graben downdropping the Botwood Group against older rocks also appears to indiscriminately cut across the F_1 fold structures. The east-west to locally north-south trending fault that downdrops the Goldson conglomerate against older rocks is also inferred to post-date the F_1 folds. Therefore, most of the faults within the area except the east-west trending fault that separates the Luscombe and New Bay Formations north of Campbellton along Rideout Point may be considered post-F_1 structures.

The attitudes and styles of all mesoscopic folds and cleavages are roughly similar. Figures 4-13 and 4-14 represents a composite lower hemisphere, equal-area stereographic plot of poles to S_0 and S_1 and F_1/L_1 lineations from all the subareas. If it is assumed that these mesoscopic structures are fabric elements of controlling macroscopic structures, then these stereographic plots imply that a series of macroscopic moderate to steeply southeast plunging, inclined to reclined, close to tight, overturned folds is found within the map area. Axial surfaces of these folds appear to strike roughly northeast-southwest and dip vertically to moderately southeastwards (Fig. 4-14).

Poles to S_1 (slaty cleavage) and F_1/L_1 lineations appear to be somewhat redistributed. The redistribution pattern and steeply plunging to reclined nature of these folds may be explained in several ways. Folds with steeply plunging axes may result from two phases of folding. If a phase of large-scale, upright, isoclinal folding of bedding about northeast-southwest trending horizontal axes with no axial surface cleavage development or only a locally developed slaty cleavage preceded a phase of folding characterized by moderately to steeply dipping axial planes with an axial surface slaty cleavage, the fold pattern given in figure 4-13 could result (see Turner and Weiss, 1963, page 142). A final phase of large-scale gentle folding could have then redistributed the S_1 and F_1/L_1 attitudes. Alternately, the F_1 folds and their axial surface slaty cleavage may represent a set of originally recumbent folds that have been refolded about a

Lower Hemisphere equal-area Stereographic Projection



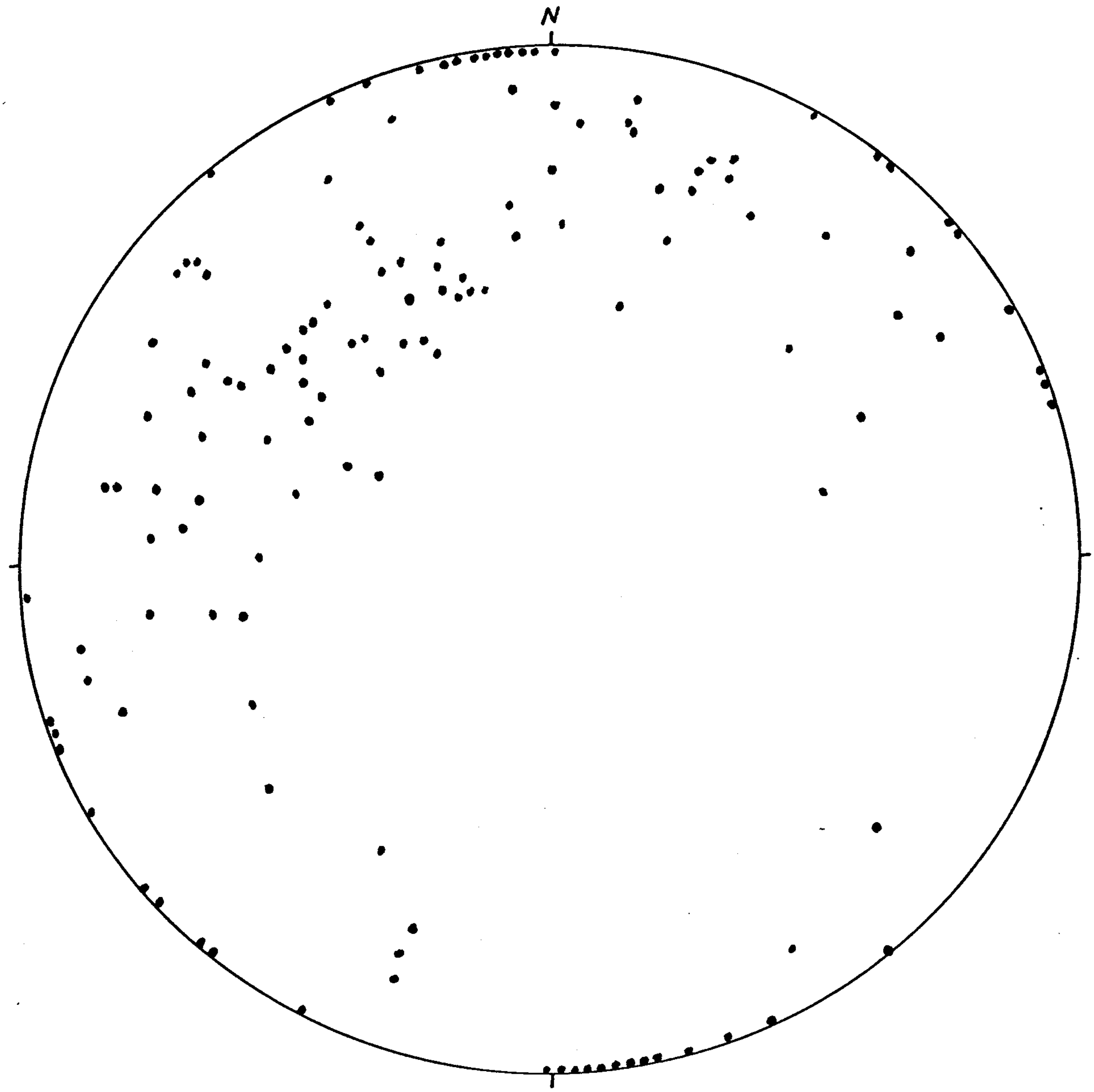
Composite of all Subareas

• poles to S_o (bedding)

+ F_1 hinges and L_1 lineations

Figure 4-13

Lower Hemisphere equal-area Stereographic Projection



• poles to S_1 (slaty cleavage)

Composite of all subareas

Figure 4-14

series of large-scale, upright, northeast-southeast trending open folds. This would also result in the slight redistribution of S_1 and F_1/L_1 attitudes. A third way to explain the observed fold pattern is that a series of originally upright folds (F_1) have been systematically tilted and redistributed by the en echelon north-northeast faults that cut through the map area. This is unlikely however since these faults are quite numerous and if they did redistribute the folds, the stereographic patterns would be expected to be much more chaotic. Therefore, it is possible that the structure in this area resulted from two phases of superimposed folding or perhaps the pattern reflects one single progressive inhomogeneous deformational event that resulted in fold culminations and depressions (Ramsay, 1962). A general pattern of northeast trending folds affects the rocks of the Exploits terrain of Notre Dame Bay (Williams, 1963; Dean, 1977; Helwig, 1967; Nelson, 1979). The southeast trending folds found within my map area appear to depart from this general northeast fold trend implying that a local second phase of folding may have affected my map area. Horne (1968) identified two phases of folding on New World Island just to the northeast of the Loon Bay/Lewisporte area. The first phase of folding was the major deformation and involved the development of originally east-west trending upright folds with a regionally penetrative axial surface slaty cleavage (Horne, 1968). These folds were then refolded about south-plunging folds with only a locally developed 'strain-slip' axial surface cleavage (Horne, 1968). Unfortunately, my area is too small to elucidate the actual series of deformation events was.

The youngest rocks folded in my area are of upper Ordovician age. The regional northeast trending folds of the Exploits terrain fold rocks as young as early Silurian in age (Dean, 1978). It is commonly assumed that folding in this region is of Devonian (Acadian) age (for example, see Williams and others, 1974). The structural history of my map area may be summarized as follows:

D_0 - soft sediment slumping and development of non-penetrative, polyclinal soft sediment folds.

D_1 - Development of moderately to steeply plunging inclined to reclined folds with an associated penetrative axial surface slaty cleavage (Acadian age). This may have involved two super-imposed phases of folding.

D_2 - Development of north-northeast/north-northwest trending conjugate strike-slip fault set and associated fracture sets of the same trend. Minor kink bands are locally seen kinking earlier S_1 slaty cleavage. Also, downdropping of the Silurian Botwood Group and Goldson conglomerate along moderately inclined to steep gravity faults.

CHAPTER V

CONCLUSIONS

Plate Tectonic Scenario

Development of the central Newfoundland ensimatic island arc (Dunnage zone of Williams, 1979) began sometime in the media or late Cambrian (Dewey and Bird, 1971; Williams, 1979) and continued to develop through the medial Ordovician until Caradocian times (Nelson and Casey, 1979). The volcanogenic manganiferous cherts of the Luscombe Formation are obviously related in some way to this late Cambrian/Ordovician Island arc system. The important question is exactly what this relationship may be. Is the Campbellton Sequence a back-arc or front-arc assemblage?

Since the Campbellton sequence is proposed to underlie the Dunnage melange (see Stratigraphic Relationships section) the origin of this sequence is intimately tied to the origin of the Dunnage melange. Dewey (1969) first suggested that the Dunnage melange formed within an accretionary prism related to a west-dipping subduction zone. Kay (1976) also believed that the Dunnage melange formed in a west-dipping subduction zone that ceased just prior to Caradocian times. The main objections to the Dunnage melange forming in an accretionary prism is that it does not contain any ophiolitic blocks or show any signs of blueschist facies metamorphism that is common to most accretionary prisms (see Ernst, 1974). Furthermore, as previously discussed the Dunnage melange appears to be an olistostrome deposit related to gravitional slumping into a trough rather than being tectonically churned up in an accretionary prism (see Hibbard and Williams, 1979).

Recently Pajari and others (1979) described the Carmanville ophiolitic melange found east of the Dunnage melange as an olistostromic deposit containing large blocks of sedimentary, volcanic and ultramafic rocks set in an argillaceous matrix that is much more intensely deformed than the Dunnage melange. The Carmanville melange contains (?)early Cambrian to Caradocian age fossils (Pajari and others, 1979) whereas the Dunnage melange contains rocks as old as middle Cambrian and is overlain by Caradocian age shales (Hibbard and others, 1977). Due to the numerous lithological and temporal similarities between both the Carmanville and Dunnage melanges Pajari and others (1979) tentatively correlated these two melange units that are separated by several kilometers of Silurian sedimentary and volcanic rocks. Horne (1969), Hibbard and Williams (1975) and Kay (1976) also suggested that these two melange units are correlative.

The Franciscan melanges of western California may have been assembled by olistostromic accumulations that slide off the inner slope into the trench floor or axes (Page, 1978; subduction zone terminology from Dickinson and Seely, 1979). The ophiolitic Carmanville melange could represent a partially preserved series of accreted wedges and their associated trench floor olistostromes that formed over a west-dipping subduction zone during the development of the central Newfoundland volcanic arc represented by volcanic rocks to the west of the Dunnage/Carmanville melanges (see Dewey and Bird, 1971). If this is the case, then the Dunnage melange may represent olistostromes fed into the forearc trough between the Carmanville accretionary prism to the east and the volcanic arc to the west (Fig. 5-1). The thin-bedded volcaniclastics of the New Bay Formation would sit on the western edge of the forearc trough. Conglomeratic turbidites represented by the conglomeratic portion of the New Bay Formation and the Riding Island greywacke appear to have cut channels into the edge of this forearc trough. Large-scale slumping of this material would result in the olistostromes of the Dunnage melange.

Grippi and Burke (1980) have described similar olistostromic deposits of the Tom Springs Formation in a Cretaceous age forearc trough of Jamaica where large andesite blocks up to 1 kilometer across float in a silty volcaniclastic matrix. Rapid uplift of the edges of this trough near volcanic centers and arc related earthquakes are suggested to have triggered these large olistostrome events into the forearc trough of Jamaica (Grippi and Burke, 1980).

If the Dunnage melange formed in a forearc trough than it logically follows that the Campbellton sequence was deposited in this same forearc trough since it is interpreted to underlie the Dunnage melange. During Cambrian times the central Newfoundland ensimatic island arc was in its incipient stages of development (Williams, 1979). Some of the early volcanism of this arc would be subsea in nature (Fig. 5-2). Intense subsea volcanism would result in much hydrothermal activity dissolving large amounts of silica and metals such as manganese and iron into the sea water leading to the deposition of the Luscombe Formation in areas adjacent to the arc such as the forearc basin. The restricted regional extent of the Luscombe Formation may reflect its deposition in an isolated basin, or perhaps similar rocks were deposited and subsequently covered by sediments of the forearc

Schematic Block Diagram of lower/middle Ordovician Tectonic Setting

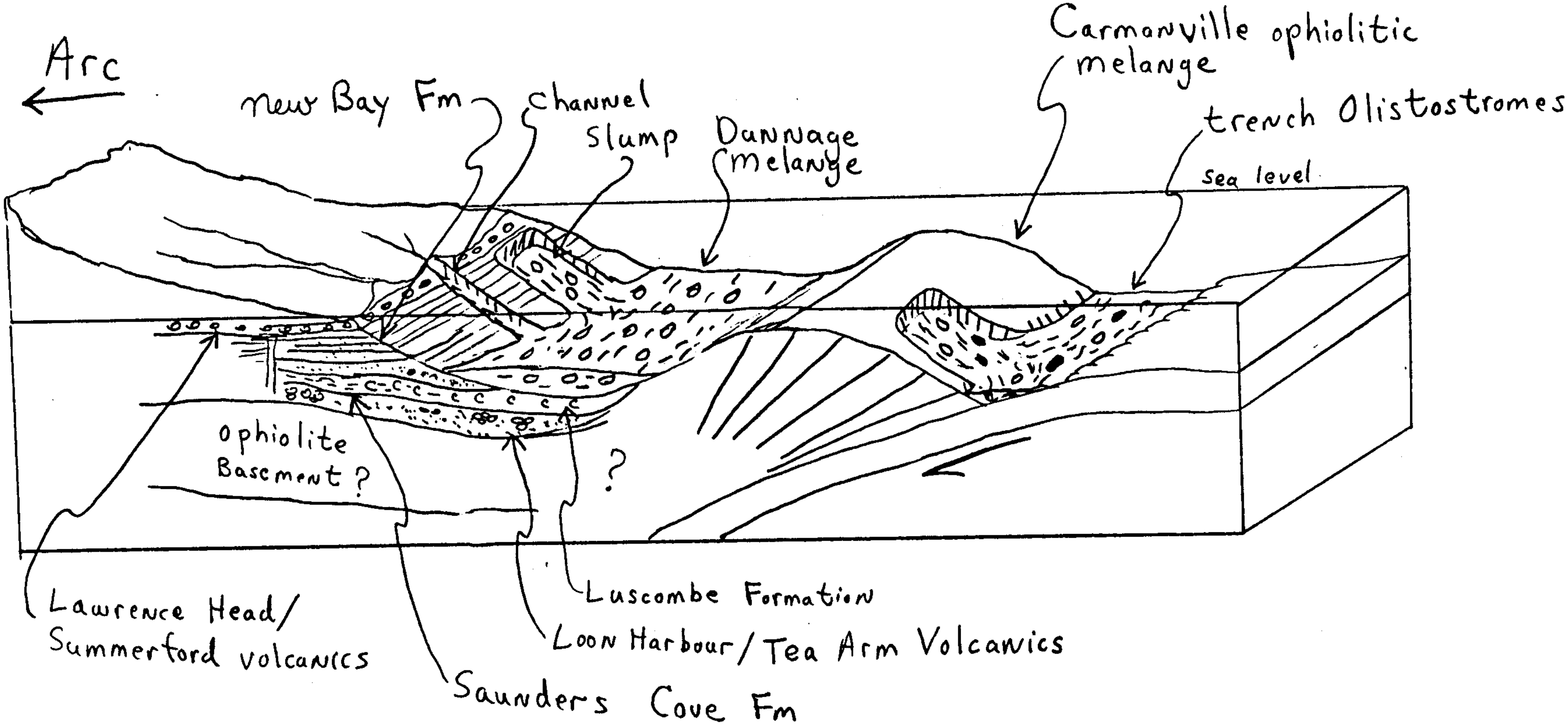


Figure 5-1

Schematic block diagram of medial/late Cambrian Tectonic Setting

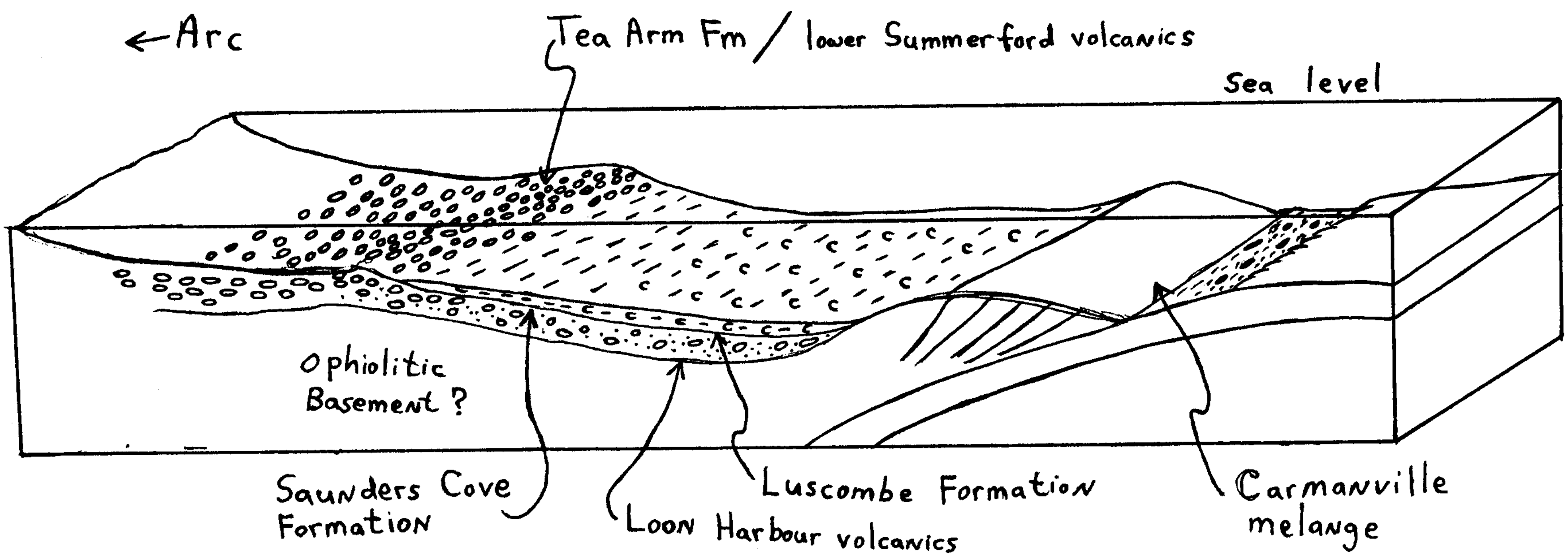


Figure 5-2

trough, such as the Dunnage melange, and are not presently exposed. A bed of manganiferous chert found within the Dunnage melange on Camel Island (see manganiferous nature of the Dunnage melange section) implies that deposition of manganiferous cherts contained but in a more limited way during deposit of the Dunnage melange. Manganese-bearing mafic to intermediate volcanic rocks extruded during the early stages of arc development could have been uplifted in the lower or middle Ordovician to provide a manganese-rich volcanic source for the New Bay Formation (see discussion on the manganiferous nature of the New Bay Formation and Dunnage melange in Lithology section).

Alternately, the Loon Harbour and Luscombe Formations may have been deposited in a back-arc marginal basin setting associated with an east-dipping subduction zone west of this area that resulted in development of the central Newfoundland Island arc (see Nelson and Casey, 1979). Intense subsea volcanism related to back-arc spreading or distal arc volcanism in middle/upper Cambrian (Corner Point, Tea Arm and Lower Summerford volcanics) would produce the needed hydrothermal activity to permeate the sea water with large amounts of silica and manganese resulting in deposition of the Luscombe Formation.

All arc volcanism ceased just prior to the Caradocian times as the Bay of Islands ophiolite was obducted onto the platform to the west (Nelson and Casey, 1979) and the Gander ophiolite underlying the Carmanville melange was obducted onto the 'Ganderland' continental mass to the east (Pajari and others, 1979). These events were the culmination of the Taconic orogeny in Newfoundland. Since opposed ophiolite obduction events occurred synchronously, it implies that a middle Ordovician period of opposed subduction must have occurred.

That is, prior to the obduction events both west and east-dipping subduction zones must have existed (Dewey and Bird, 1971). Thus, the Loon Harbour and Luscombe Formations could have been originally deposited in a middle/upper Cambrian back-arc basin that began to be subducted westwards beneath the central Newfoundland Island arc in the lower/middle Ordovician resulting in the situation illustrated in Figure 5-1. Strong (1977) and Nelson and Casey (1979) have discussed the evolution of an east-dipping subduction zone beneath central Newfoundland whereas Williams (1979) has advocated a west-dipping subduction zone. Since both of these theories could be correct, the real question is which subduction zone resulted from the closure of a 'large' ocean and which subduction zone formed by the closure of a marginal basin and how synchronous were these opposed subduction zones.

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