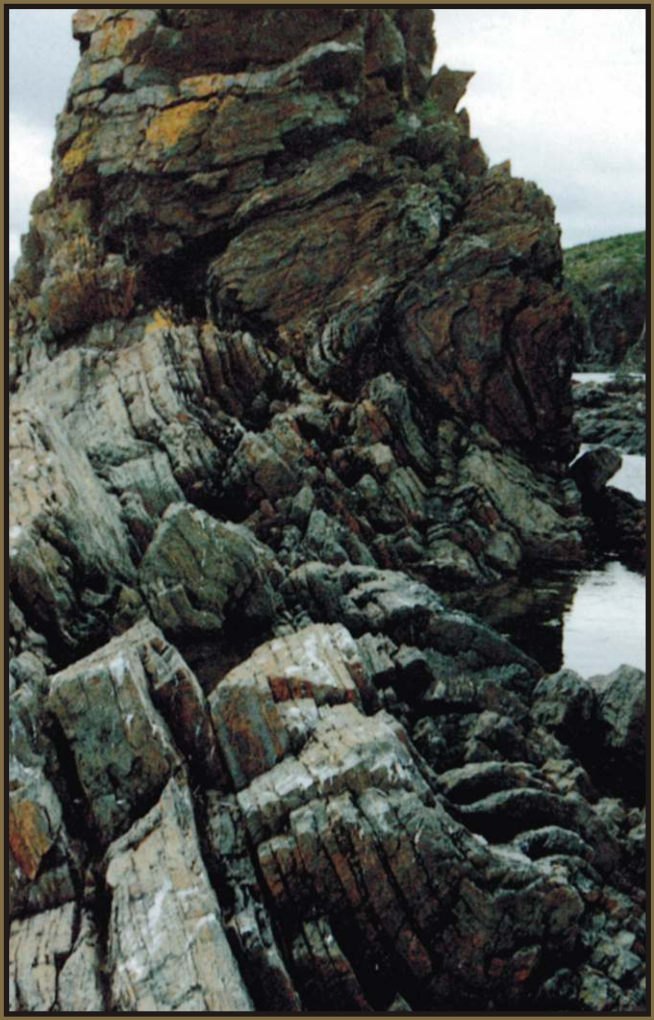
GEOLOGY OF THE CENTRAL NOTRE DAME BAY REGION (PARTS OF NTS AREAS 2E/3,6,11), NORTHEASTERN NEWFOUNDLAND

Brian H. O'Brien, P.Geo.

Report 03-03

St. John's, Newfoundland
2003



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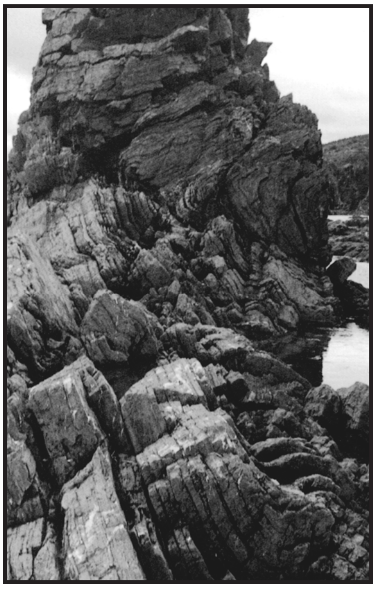

Cover

Sea stack of Ordovician Wild Bight Group cherts showing an antiformal subrecumbent syncline (and associated limb thrust) refolded by upward-facing minor folds having an opposing sense of fold asymmetry. Beds in the foreground are right-side-up; beds at the top of the sea stack are overturned. Situated on Red Island, off Alcock Island, near Leading Tickles, Notre Dame Bay, Newfoundland.

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EDITING, GRAPHIC DESIGN AND CARTOGRAPHY

Senior Geologist	B.F. KEAN
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Copy Editor	D.G. WALSH
Graphic Design, Layout and Typesetting	D. DOWNEY J. ROONEY B. STRICKLAND
Cartography	D. LEONARD A. PALTANAVAGE T. SEARS

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Email: pub@zeppo.geosurv.gov.nf.ca
Website: http://www.geosurv.gov.nf.ca

Author's Address: Brian H. O'Brien, P.Geo.
Geological Survey
Department of Mines and Energy
P.O. Box 8700
St. John's, NL
Canada
A1B 4J6

NOTE

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ABSTRACT

The central Notre Dame Bay region comprises part of the type area of the oceanic Dunnage Zone of the Canadian Appalachians. Two lower Paleozoic subzones are represented; the Notre Dame Subzone in the northwest and the Exploits Subzone in the southeast. These are tectonically separated by the Red Indian Line. The lower–middle Paleozoic overstep sequence of the Exploits Subzone is particularly well developed in the central Notre Dame Bay region.

The lithostratigraphy of the Exploits Group and easternmost Wild Bight Group (Exploits Subzone) is systematically presented, as is the biostratigraphy of the overstep sequence units. In the area surveyed, the latter include the Lawrence Harbour Formation, the Shoal Arm Formation, the Luscombe Formation, as well as the succeeding formations of the Badger Group and the Botwood Group. Lithostratigraphic units, such as the Moretons Harbour Group, the Cottrells Cove Group and the Boones Point Complex of the Red Indian Line structural zone, have been historically included in the local lithotectonic sequence of the Notre Dame Subzone.

The structural development of the central Notre Dame Bay region was controlled by a superimposed sequence of four regional deformations. Tectonic structures, which are treated by reference to specific field examples in section and in plan, were produced on various scales during each deformation. The early to late Paleozoic history of the pretectonic plutonic rocks of the South lake and Philips Head igneous complexes is outlined. These magmatic rocks are contrasted with the major posttectonic intrusions in parts of the Loon Bay, Hodges Hill and Mount Peyton batholiths.

In the central Notre Dame Bay region, various stratified and intrusive rocks from a wide range of lithologic units are observed to be altered. Certain lower Paleozoic units have potential for volcanogenic base-metal deposits or sediment-hosted precious and base-metal deposits. Many middle Paleozoic and older units have potential for epigenetic precious-metal mineralization.

INTRODUCTION

LOCATION AND ACCESS

Most of the geological mapping for this report was carried out in central Notre Dame Bay within National Topographic System (NTS) map areas 2E/3,6 and 11 between 1989 and 1993 (Figure 1). Reconnaissance studies were made in adjoining areas as the rock groups discussed herein are widespread throughout north-central Newfoundland. The region surveyed includes the western Bay of Exploits and the area south to Notre Dame Junction, the Fortune Harbour Peninsula and the fjords of New Bay, and the lower reaches of the New Bay and Exploits rivers south of Point Leamington and east of Bishop Falls. Boundaries of the map area are latitude 49°33' in the north, latitude 49°1' in the south, longitude 55°29' in the west and longitude 55°1' in the east.

Access to the study area is largely on marine waters by open boat from villages and towns serviced by all- weather roads. Canoe traverses were staged from forest access roads and allied tracks, themselves accessible by off-road and all-terrain vehicles. All parts of the map area can be reached by a single two-person traverse team without the aid of aircraft.

Rock exposures along coastline, lakes, rivers and roads were examined in detail; non-systematic inland traverses were conducted where necessary to gather facts to aid interpretation. While some regions were covered using 1:50 000 aerial photographs and data was collected on this scale, observation density in other areas led to 1:10 000 scale map coverage.

PREVIOUS GEOLOGICAL WORK

Regions mapped by other workers within and adjacent to the study area are shown in Figure 2. Most of these works have preceded the analyses carried out for this report, although more recent geological surveys and pertinent map-based theses are also illustrated. They represent over five decades of geological mapping in this economically prospective and relatively well-studied part of Newfoundland.

The Notre Dame Bay region underwent an early phase of mineral exploration during the nineteenth-century copper boom in North America and western Europe. However, it first became a focus of wider geological interest in the earliest twentieth century following isolated discoveries of early Paleozoic fossils in volcanosedimentary rock sequences. These occurrences prompted a closer regional correlation with the eugeosynclinal strata of Appalachian North America and Caledonian Europe than with the fossiliferous miogeosynclinal strata of western Newfoundland or the strata of the St. Lawrence Platform on the Canadian mainland and the continental interior of the United States. Geological mapping in central Notre Dame Bay was initially aimed at placing the known fossil localities in a lithostratigraphic framework. Early investigations attempted to recognize age equivalent or correlative map units in sparsely fossiliferous lithotectonic sequences dominated by wacke and basalt. Thus began the debate over the definition of the Ordovician Exploits Group and, subsequently, the Exploits Subzone of the Dunnage Zone (*see* Williams, 1995b for summary).

G.R. Heyl (1936) surveyed the east-central coast of Notre Dame Bay (Figure 2). Based on the presence of *in situ* Normanskill (Llandeilo–Caradoc) and reworked Chazy (Llanvirn–Llandeilo) fossils, he assumed that the stratified rocks were all mid-Ordovician and assigned them to his Exploits Series (Table 1). Heyl (1936) did acknowledge the existence of Late Ordovician and Early Silurian fossil-bearing wackes in Notre Dame Bay, although he restricted their outcrop to the eastern part of New World Island. However, as originally defined, the Exploits Series contained several unrelated rock groups and formations which were either improperly correlated or assigned to an incorrect stratigraphic position relative to the regional black shale marker that had yielded the Normanskill fauna (the Lawrence Harbour shale). Moreover, different ages of melange were grouped together or inappropriately affiliated with the wrong tracts of unbroken formation. Accordingly, the term Exploits Series fell into disuse (*see* Chart of Abandoned Terms in Table 1).

G.H. Espenshade (1937) demonstrated the lithological continuity and tectonic linkage of certain rocks in the west-central coastal region of Notre Dame Bay with those surveyed by Heyl (1936). These were included in his now-supplanted Badger Bay Series (a partial correlative of the Exploits Series). Significantly, to the northwest of the Badger Bay Series, Espenshade (1937) recognized a fundamentally different sequence of probable age- equivalent and older Ordovician rocks that he termed the Pilleys Series. These included the limestone-bearing volcanosedimentary strata of his Cutwell Group and what Esphenshade (*op. cit.*) considered were the younger overlying strata of the Lushs Bight Group farther north. However, his Roberts Arm volcanics and Crescent Lake shales, which were mapped to the southeast of Pilleys Series rocks and stated to be directly faulted against them, were incorrectly assigned to the upper part of the Badger Bay Series. Moreover, they were stated to be in stratigraphic conformity with rocks comprising the lower part of the Badger Bay Series (i.e., his Wild Bight–Shoal Arm–Gull Island succession). The deformed terrestrial sedimentary and volcanic rocks of the Springdale Group were purported to be either Silurian or Carboniferous in age and were interpreted to lie above the Badger Bay Series with angular unconformity (Espenshade, 1937).

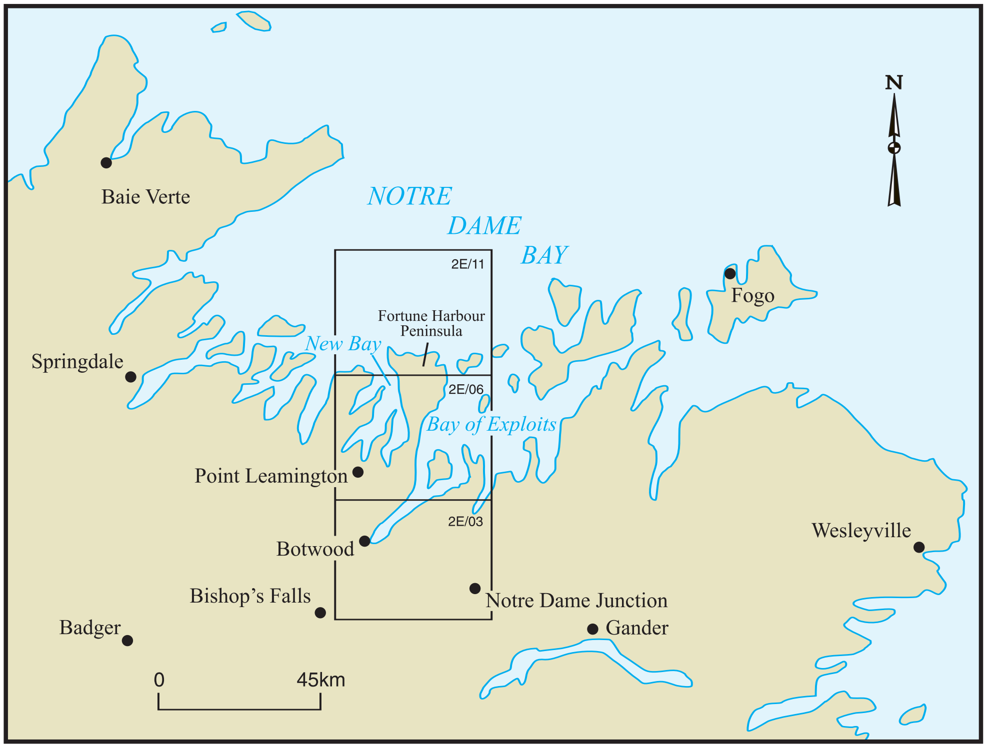


Figure 1. *Geographical location of NTS map sheets 2E/3, 6 and 11 in central Notre Dame Bay, northern Newfoundland.*

J.J. Hayes mapped a large part of south-central Notre Dame Bay (Figure 2) around the time that Newfoundland entered the Canadian confederation and he originally replaced the Badger Bay Series with the term Exploits Group (Hayes, 1951). However, like Espenshade (1937), he included the lithotectonic sequence from Wild Bight Formation to Roberts Arm Volcanics in the Exploits Group. Hayes also concurred with Heyl by incorrectly placing the Gull Island member of his Sansom Formation stratigraphically beneath the Crescent Lake Formation and, most importantly, by correlating the Crescent Lake shale with the Normanskill black shale at Lawrence Harbour. As a result, the Roberts Arm Volcanics were held to be somewhat younger than the purported mid-Ordovician Crescent Lake shale of the Exploits Group. Hayes' (1951) observations on the disposition of lithological units near the western margin of his Middle Ordovician (?) Exploits Group are consistent with the repetition of said units by large-scale polyphase folding and regional fault imbrication.

T.O.H. Patrick (1956) systematically mapped the region to the immediate east of the area surveyed for this report (Figure 2). At that time, debate remained focused on the definition of the Exploits Group. As with previous workers, he considered this rock group to be mostly mid-Ordovician in age and he embodied the turbidites of his Sansom Formation in the basal part of the group. He also assigned rocks now included in the Dunnage Melange and the Summerford Group to his Exploits Group. Patrick (1956) placed terrestrial rocks thought to be correlative with the Springdale Group stratigraphically above his shallow-marine Indian Islands Group. However, the terrestrial strata of his Farewell Group were situated stratigraphically below a marine succession comprised of his Sansom Formation and the Indian Islands Group (*see* Chart of Abandoned Terms in Table 1).

In the sixties, Harold Williams produced a 1:253 440 geological map of the NTS 2E district as part of a regional synthesis of the Notre Dame Bay area (Williams, 1962; Williams, 1963a, 1963b; Williams, 1972). Fossil-bearing Middle and Upper Ordovician sedimentary units, an underlying volcanic unit and various melange units were included in the Exploits Group. These were positioned stratigraphically above the Middle Ordovician and older Wild Bight Formation, which Williams (1963) elevated to group status. At this time, Williams (*op. cit.*) also reassigned Patrick's Sansom Formation and correlative rocks to the upper part of the Exploits Group, placed the structurally overlying Crescent Lake Formation and Roberts Arm Volcanics in the newly named Roberts Arm Group, and considered the mutual boundary of the redefined Exploits and Roberts Arm groups to be, everywhere, a faulted contact. Williams (1962) did not make a lithostratigraphic distinction between the rocks of the Exploits and Summerford groups and, following Patrick (1956), he included the Dunnage Melange within the Exploits Group. In essence, the surveys carried out by him and earlier workers defined what would later become accepted as the Ordovician stratigraphic template of the Exploits Subzone of the Dunnage Zone (Williams *et al.*, 1988).

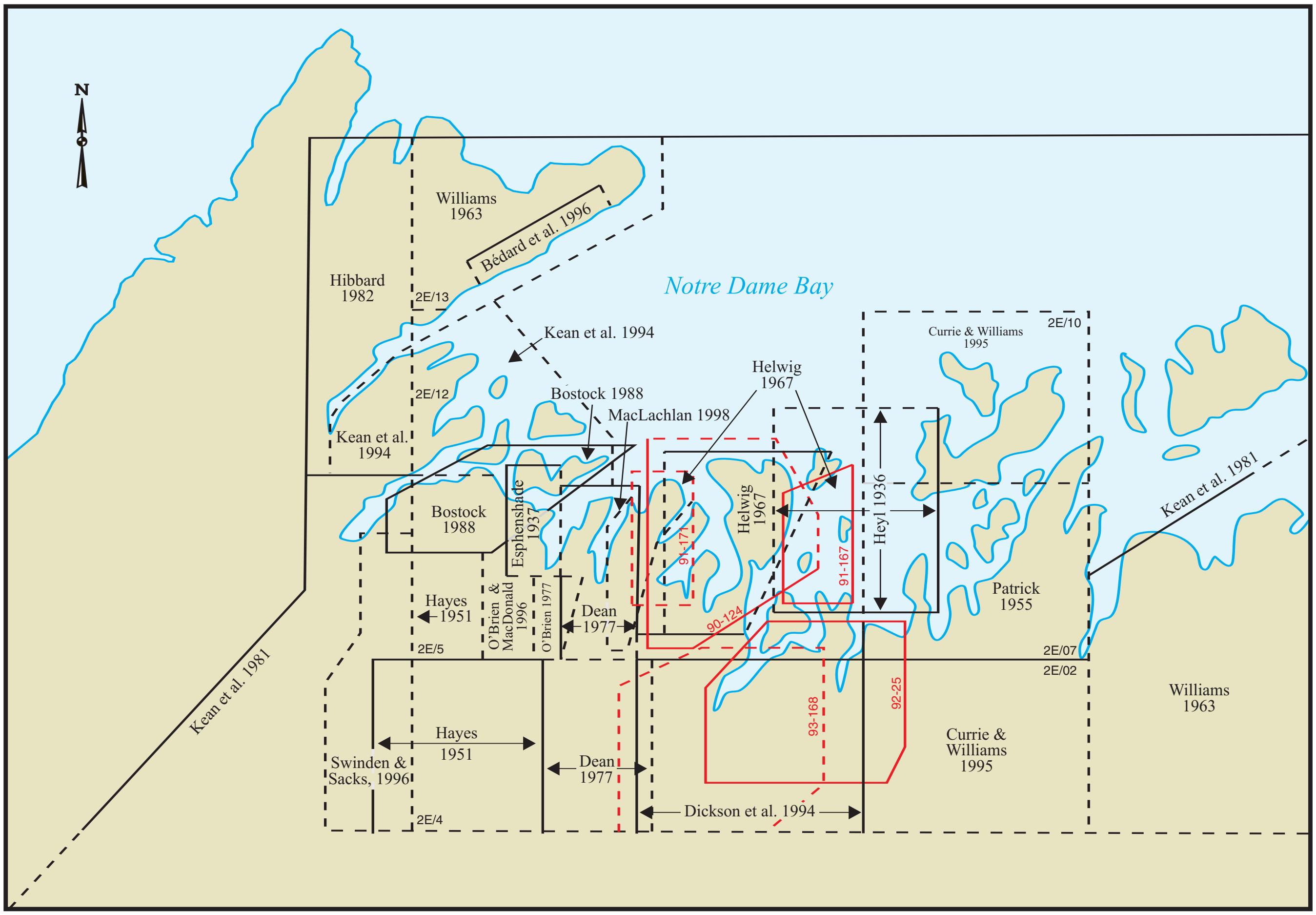


Figure 2. *Geological index map of Notre Dame Bay showing most of the previous regional mapping studies in the NTS 2E district of northern Newfoundland. Index citations are to the author of the map and its year of publication; the full bibliographic reference accompanies this volume. Red outlines and their numbers relate to maps producted by the writer during the course of this study.*

J.A. Helwig mapped the northern part of the study area in detail (Figure 2). In 1969, he revised the Exploits Group and formally named six of the conformable internal formations (Table 1). He recognized a 2 km thick sequence of interbedded wacke and argillite with minor conglomerate lenticles and demonstrated that the succession was mid and late Ordovician in age. Helwig termed this late Caradoc and Ashgill lithostratigraphic unit the Point Leamington Greywacke, and placed it stratigraphically above the Llandeilo–Caradoc (Normanskill) black shales recognized by the earliest workers. The graptolitic beds in gradational contact with his basal Point Leamington Greywacke resided in an unnamed late Caradoc unit which, on new fossil evidence, was slightly younger than his Middle Ordovician Lawrence Harbour Shale. Helwig (1967) was one of the first workers in Notre Dame Bay to caution that younger parts of the regional black shale formation were commonly overthrust by older parts of the same formation. However, following tradition, he included the Point Leamington Greywacke and redefined Lawrence Harbour Shale in his uppermost Exploits Group.

By way of contrast, J.A. Helwig excluded purported Central Notre Dame Bay correlatives of the Early Silurian (Llandovery) Goldson Formation from his upper Exploits Group (Helwig, 1969). He did this knowing that a red marine conglomerate of the Goldson Formation locally unconformably overlies mid-Ordovician (Llanvirn–Llandeilo) pillow lava some 75 km to the east on New World Island and that, less than 50 km to the west of the study area on Pilleys Island, a red terrestrial conglomerate is exposed above the angular unconformity at the base of the Silurian Springdale Group (Figures 3 and 4). He also excluded the Late Ordovician Sansom Formation (partial biostratigraphic equivalent of the Point Leamington Greywacke) from his Exploits Group and restricted this sub-Goldson formation to the New World Island area. Helwig (*op. cit.*) thought that a Late Ordovician–Early Silurian hiatus was present in the

Exploits Group in the Bay of Exploits near Upper Black Island (*see* Map 2001-41; click here), where a major intraformational disconformity marked the region separating the Point Leamington Greywacke from the Sansom Formation[1]. He also interpreted a mid Ordovician hiatus in the Wild Bight Group in the Seal Bay–Badger Bay area of west-central Notre Dame Bay. Moreover, Helwig postulated that, beneath the localized disconformity within the Wild Bight Group, lay the correlatives of his lower and middle Exploits Group.

Rocks underlying the northernmost part of the Fortune Harbour peninsula (Figure 2) were assigned to Helwig's (1967) Lushs Bight Group (Figure 3; Table 1). Although correlating the unit on the peninsula with the type area farther west, he departed from Espenshade's original definition by including probable equivalents of the Roberts Arm volcanics and Crescent Lake shales in a southern volcanosedimentary tract. Helwig (1967) interpreted this southern tract as the stratigraphically lowest subunit of his Lushs Bight Group. A younger subunit to the north was thought to be mostly composed of strata from the original Lushs Bight Group of the Pilleys Series but also included strata from the lower volcanic-dominant part of the Cutwell Group of the Pilleys Series.

G.S. Horne and J.A. Helwig made a regional revision of the Ordovician stratigraphy throughout most of the area shown in Figure 2 (Horne and Helwig, 1969). They emphasized that Ordovician volcanic, plutonic and sedimentary rocks included in Espenshade's (1937) Pilleys Series, Williams' (1963) Roberts Arm Group and Helwig's (1967) Lushs Bight Group occur north of the Lukes Arm Fault Zone (Figure 3) and considered that these units faced northward in the stratigraphic sense. Furthermore, such rocks were thought not to be directly correlatable with the sparsely fossiliferous Ordovician volcanic and Ordovician–Silurian sedimentary rocks that lay to the south of this melange-bearing fault zone, though some rock types were stated to be remarkably similar. Horne and Helwig (1969) were the first workers to purport the existence of both early and middle Ordovician volcanic formations in the Exploits, Wild Bight and Summerford groups, and located all these volcanic rock-bearing units stratigraphically beneath a regionally-developed (but locally named) unit of Caradoc black shale (Table 1). In general, they concurred with Williams (1963) in placing the Late Ordovician and Early Silurian turbidites above the late Middle Ordovician black shale and older volcanic rocks, and in regionally separating the younger siliciclastic and calcareous flysch from Notre Dame Bay's Early and mid-Ordovician volcanogenic turbidites.

In 1977, P.L. Dean produced a series of metallogenic maps of north-central Newfoundland. In these publications, he consistently compiled the regional geology of most of the area shown in Figure 2 at 1:50 000 scale for the first time. Most of his original work was focused on the regional setting of Ordovician base-metal deposits and showings in the Wild Bight, Exploits and Summerford groups, and on how these mineralized rocks related to prospective but lesser known volcanic sequences in adjacent rock groups to the north, south and west. Dean (1977) separated the Wild Bight Group into five conformable formations (with variable mineral potential) and originally proposed an internal stratigraphy for what he considered were its Lower and Middle Ordovician rocks. As with previous workers, he correlated parts of the Wild Bight Group with the Exploits and Summerford groups, but treated each as a lithostratigraphically distinct unit[2].

North of the Exploits Group and the Point Leamington Formation, Dean (1977, 1978) recognized a new volcanosedimentary unit which he called the Cottrells Cove Group (Figures 3 and 4; Table 1). He originally defined the basal Moores Cove and overlying Fortune Harbour formations of his northward-facing Cottrells Cove Group and considered that these formations were the general equivalents of (and thus superseded) the southern or lower units of Helwig's (1967) Lushs Bight Group. Dean correlated certain rocks north of the Cottrells Cove Group with the Western Head Formation of the Moretons Harbour Group, the type section of which is exposed near New World Island and Twillingate in eastern Notre Dame Bay (Williams and Currie, 1995; Swinden, 1996; Table 1). He stated that his Cottrells Cove Group was everywhere in fault contact with southward-facing rocks in the Moretons Harbour Group on the Fortune Harbour peninsula. In reassigning the upper (northern) units of Helwig's Lushs Bight Group to this part of the Moretons Harbour Group, the early Ordovician (?) Western Head correlatives were implied to be younger than the Cambrian Sleepy Cove sequence of Twillingate but older than the Cottrells Cove Group to the immediate south (Table 1). Dean proposed an early Silurian age for Cottrells Cove volcanosedimentary strata, as he believed that their boundary with the Late Ordovician Point Leamington Greywacke was primarily stratigraphical in nature.

Dean and Strong (1977) were some of the earliest workers in Notre Dame Bay to propose a Silurian phase of regional thrusting in which the tectonic movement direction was generally from the north toward the south. They interpreted the subvertical structure at the southern boundary of the Moretons Harbour Group (the Chanceport Fault in Fig-

1 *Note:* the Point Leamington Formation was formally erected by S.H. Williams (1991) and H. Williams (1995d) informally included the Point Leamington and Sansom formations in the newly erected Badger Group.

2 *Note:* paleomagnetic studies on Exploits Group pillow lava by Todaro *et al*. (1996) and on Summerford Group pillow lava by van der Voo *et al*. (1991) indicate that their eruptive sites may have been separated by 19° paleolatitude or about 2000 km.

ure 3) to have originally formed as a north-dipping thrust, which was responsible for placing the Ordovician Western Head rocks above the purportedly Silurian rocks of the Cottrells Cove Group. However, during a subsequent regional folding event, the Chanceport thrust was thought to have been steepened and rotated into northwest- and northeast-trending segments[3].

The Lukes Arm Fault Zone, as defined by Horne and Helwig (1969), included the Chanceport Fault and various other fault strands. It separated the Wild Bight and Roberts Arm groups in the west, the Exploits and Cottrells Cove groups in the centre, and the Summerford and Chanceport groups in the east of the Notre Dame Bay region (Figure 3)[4].

Within the western part of the Lukes Arm Fault Zone, P.L. Dean identified and erected the Late Ordovician–Early Silurian Sops Head Complex and included within it various chaotically deformed sedimentary rocks and adjacent unbroken formations of volcanic and volcaniclastic strata. He assigned this melange-bearing lithostratigraphic unit to the Roberts Arm Group and positioned it immediately beneath Williams' (1963) Crescent Lake Formation. Thus, in places, the block-in-matrix melange unit lay structurally below a northwest-dipping succession of the Roberts Arm Group but above a variety of mid and Late Ordovician units in Williams' (1963) Wild Bight and Exploits groups. In regionally correlating the Roberts Arm Group with the Cottrells Cove Group, Dean (1977, 1978) generally equated the Sops Head and Boones Point melange complexes (*see also* Nelson, 1981), the Crescent Lake and Moores Cove formations, and at least parts of the Roberts Arm Volcanics and the Fortune Harbour Formation. P.L. Dean also postulated that his Silurian Roberts Arm and Cottrells Cove groups had correlatives in the volcanic and sedimentary rocks of his Frozen Ocean Group, which lay to the south of the Ordovician Wild Bight Group (Figure 3).

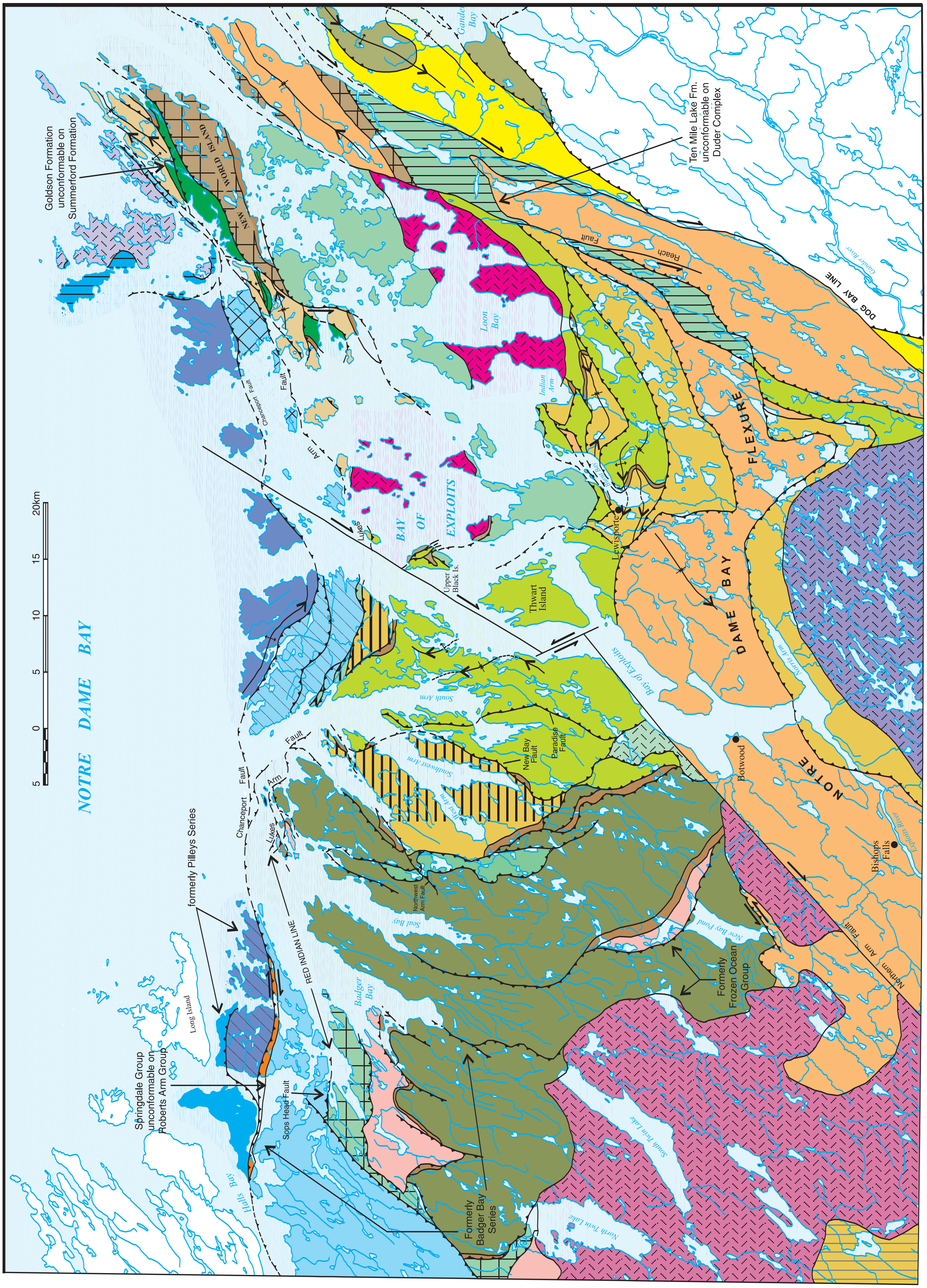
B.F. Kean and coworkers carried out systematic regional mapping of the interior of north-central Newfoundland throughout the seventies and early eighties, emphasizing the numerous base-metal-bearing volcanic units. By placing the top of the Exploits Group below the Lawrence Harbour shale, it was possible for Kean *et al*. (1981) to correlate various parts of the Exploits, Wild Bight, Summerford and Victoria Lake groups (Table 1). Integration with previous surveys of the coastal region of Notre Dame Bay resulted in a compilation map of the Central Volcanic Belt (Figure 2), which was the first regional geological map of the northern Newfoundland portion of Williams' (1979) Dunnage Zone. Kean *et al*.'s (1981) map portrayed a major tectonic break which was well illustrated by the distribution, age and origin of the lithostratigraphic units depicted. Extending over 300 km from Red Indian Lake to Twillingate and including the Lukes Arm Fault Zone, it was the first representation of what eventually became defined as the Red Indian Line (Williams *et al*., 1988). This fundamental structure separated the Dunnage Zone into two subzones, the Notre Dame and Exploits, to the west and east, respectively. These subzones were later revised and extended throughout the Canadian Appalachians (Williams, 1995). To most workers, the Red Indian Line currently ranks as the composite Laurentia–composite Gondwana suture within the Dunnage Zone (e.g., Colman-Sadd *et al*., 1992a; Tables 1 and 2).

In 1988, H.H. Bostock released a detailed 1:50 000 scale geological map of the northern part of the Roberts Arm Group and adjacent lithostratigraphic units (Figure 2). He considered the sedimentary rocks of the basal Crescent Lake Formation of his Middle Ordovician Roberts Arm Group to have once been in depositional contact with the younger tholeiitic volcanic rocks of his overlying Crescent "terrane". In doing so, he accepted Dean's stratigraphic analysis of the lower Roberts Arm Group but not his Silurian age assignment. Moreover, Bostock (1988) concluded that these undated cupriferous volcanic and sedimentary rocks (now commonly included in the Notre Dame Subzone) were structurally overlain by higher allochthons or "terranes" which contained more abundant felsic flows and pyroclastic rocks. He interpreted the upper thrust slices to represent younger parts of the Roberts Arm Group. These isotopically dated, calc-alkaline volcanic rocks were stated to be typical of the strata found near the past-producing, polymetallic base-metal mines in the Cottrells Cove–Roberts Arm–Buchans belt.

Bostock (1988) concurred with Nelson (1981) and thought that mid Ordovician basalt and fossiliferous limestone blocks in the melange tracts of the Sops Head Complex were derived from the Crescent tholeiites of the Roberts Arm Group, which was thrust southeastward in the Late Ordovician on a late Taconian or younger structure (Table 2). The Exploits Subzone footwall sequence postulated to lie immediately beneath this advancing overthrust sheet included synorogenic soft-sediments assigned to the Late Ordovician Sansom Formation. However, it also included pre-thrusting strata belonging to the Shoal Arm Formation, a unit which is age-equivalent and younger than the dated melange blocks from the Sops Head Complex. Bostock purported that the oldest rocks of the Roberts Arm

[3] *Note:* Lafrance (1989) published a more recent adaptation and refinement of the concept of an early, north-dipping, south-verging fold-and-thrust belt as a precursor to tectonic melange development near the Lukes Arm Fault at the southern margin of the Cottrells Cove Group.

[4] *Note:* a modern definition of the 2 km wide Lukes Arm–Sops Head (LASH) fault was provided by Blewett (1989), who considered it to be coincident with the Red Indian Line.


NOTRE DAME BAY
5 0 5 10 15 20km
Goldson Formation unconformable on Summerford Formation
NEW WORLD ISLAND
Ten Mile Lake Fm. unconformable on Duder Complex
Gander Bay
DOG BAY LINE
Reach Fault
Loon Bay
Indian Arm
Chanceport Fault
Lukes Arm Fault
BAY OF EXPLOITS
Lewisporte
Upper Black Is.
Thwart Island
NOTRE DAME BAY FLEXURE
Bay of Exploits
Botwood
Bishops Falls
Exploits River
South Arm
Southwest Arm
New Bay Fault
Paradise Fault
formerly Pilleys Series
RED INDIAN LINE
Seal Bay
Northwest Arm Fault
New Bay Pond
Formerly Frozen Ocean Group
Northern Arm Fault
Long Island
Badger Bay
Springdale Group unconformable on Roberts Arm Group
Sops Head Fault
Halls Bay
Formerly Badger Bay Series
South Twin Lake
North Twin Lake

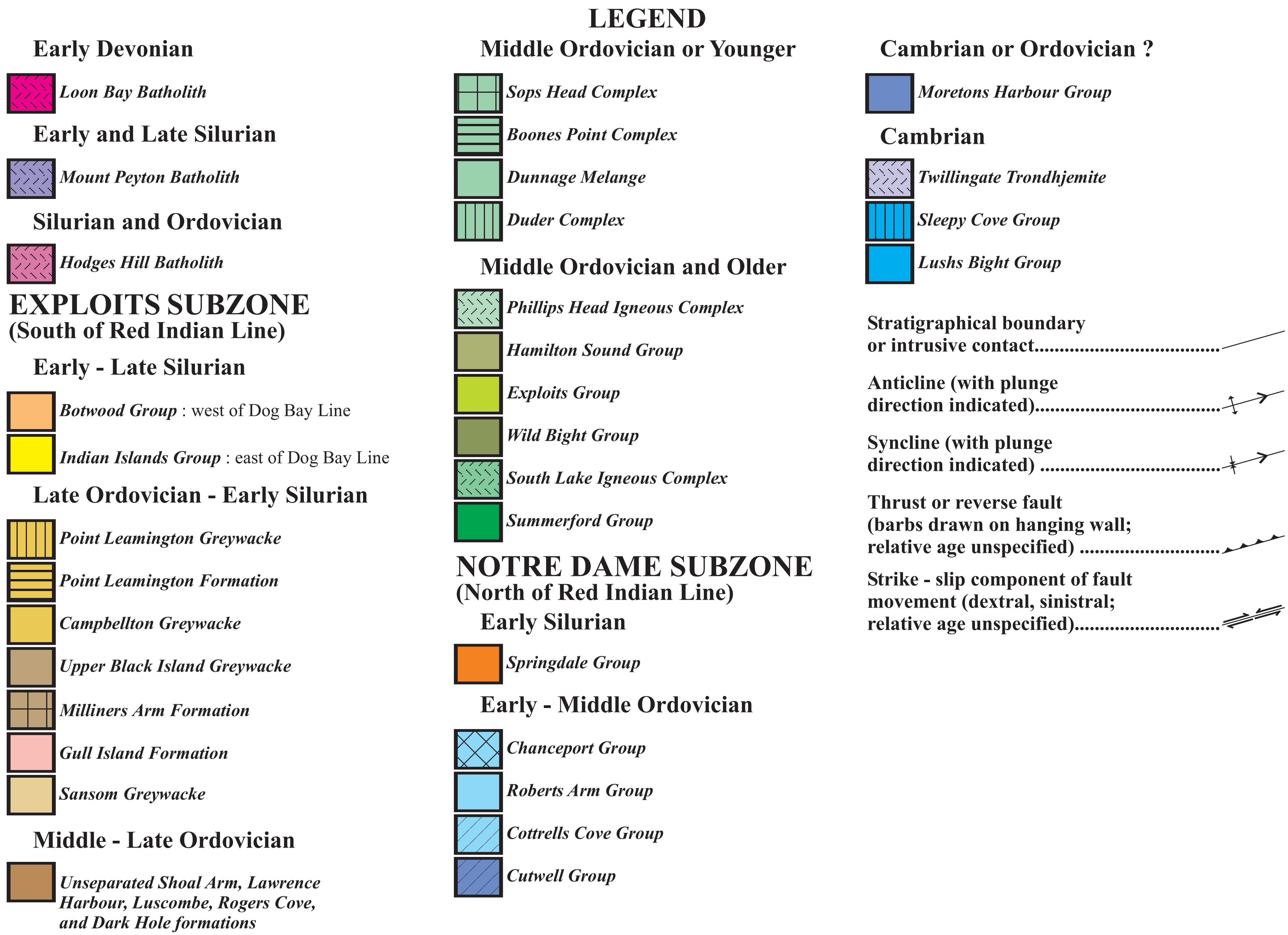


Figure 3. *Regional geological map of the Dunnage Zone between the Red Indian Line and the Dog Bay Line, illustrating the disposition of early and middle Paleozoic rocks in this part of the Central Mobile Belt of northeastern Newfoundland (NTS 2E district). The major rock units of this report are located in the central region of Notre Dame Bay. A larger map, compiled at 1:250 000 scale and colour-coded to match the Table of Formations, is located in the back pocket of this publication.* [click here for Table of Formations]

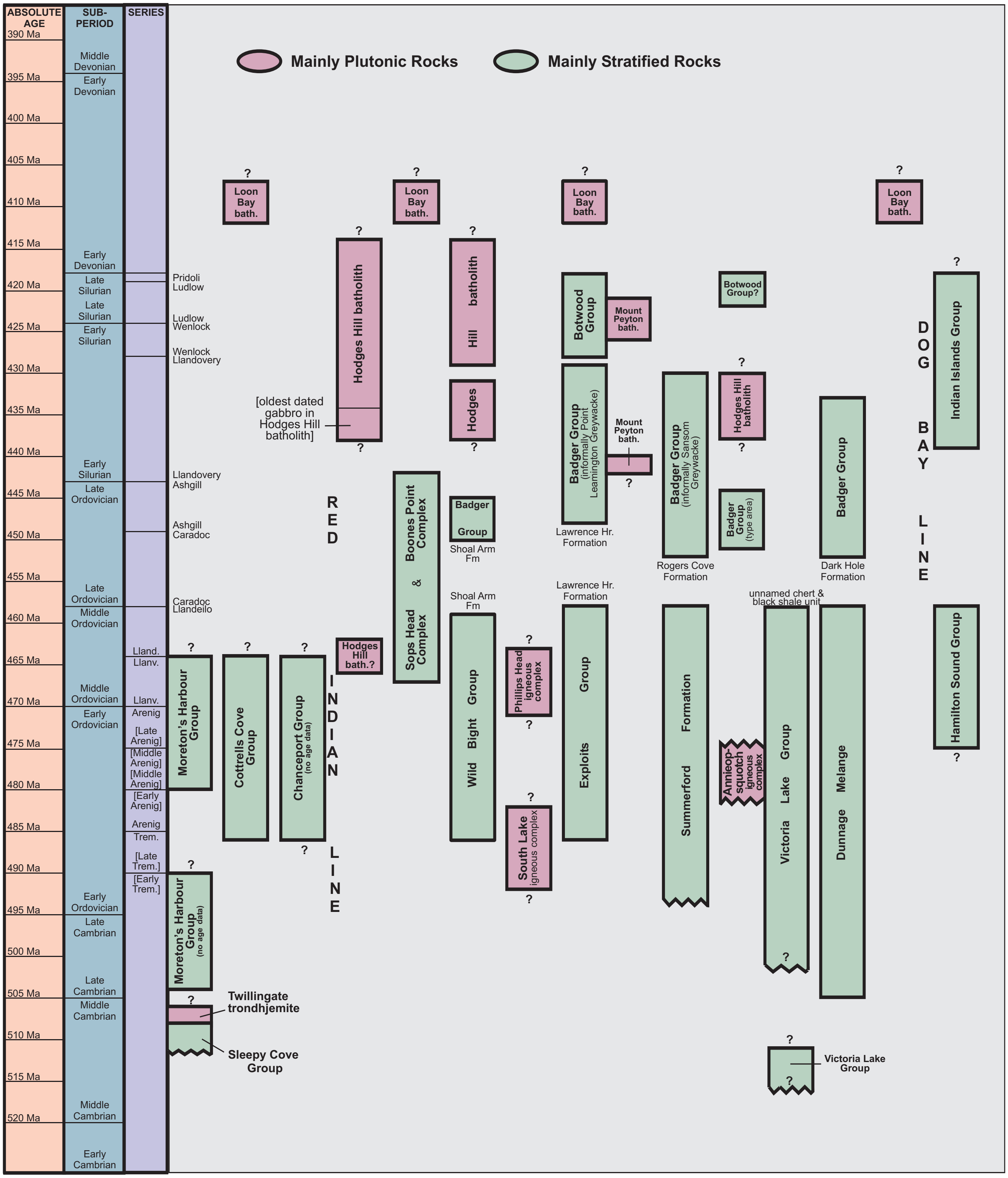


Figure 4. *Simplified table of formations highlighting the ages of the major stratified rock units and the major plutonic rock units in the Dunnage Zone of northeastern Newfoundland. Abridged from Table 1.*

Group (the Crescent Lake Formation) were also situated in the footwall plate of the regional overthrust sheet and that the hanging-wall sequence of the Roberts Arm Group contained calc-alkaline volcanic rocks that probably became younger to the northwest[5].

While confirming the unconformable relationship reported between the Silurian Springdale Group and the Ordovician Roberts Arm Group, Bostock inferred that the deformed terrestrial strata of the Springdale Group had also once unconformably overlain the marine olistostromes of the Sops Head Complex. More recently, in the central part of the Exploits Subzone, Currie (1995b) and Currie and Williams (1995) deduced that Silurian redbeds of the Ten Mile Lake Formation of the Botwood Group unconformably overlay tectonized melange belts of similar age and origin to those seen along the Red Indian Line (Figure 3). Confirming Patrick's earlier observations, these authors stated that the terrestrial deposits of the Ten Mile Lake Formation are themselves conformably overlain by turbidites of mid-late Silurian age (marine strata once referred to as the Horwood Formation of the Indian Islands Group). If correct, this would imply local emergence of the block-in matrix melange of the Duder Complex in early Silurian time near the Dog Bay Line (Williams *et al*, 1993) at the same time that the block-in-matrix melange of the Sops Head Complex became emergent near the Red Indian Line.

During the current decade, geological mapping in the vicinity of the study area (Figure 2) has focused on Notre Dame Subzone and Exploits Subzone strata adjacent to the Red Indian Line. Rocks surveyed include the mineralized Cambro-Ordovician Lushs Bight Group (cf. Szybinski, 1995) and adjacent early-mid Ordovician volcanic sequences (Kean *et al*., 1995), the terrestrial Silurian Botwood Group and adjacent Late Ordovician–Early Silurian turbidite sequences (Dickson *et al*., 1994), the marine Siluro-Devonian Indian Islands Group and adjacent mid and late Ordovician melange complexes (Currie, 1995c; Currie and Williams, 1995), the Early and mid Ordovician sequences of the Roberts Arm Group and adjacent igneous and metamorphic complexes (Swinden and Sacks, 1996; Dickson, 2000), the Betts Cove Ophiolite and adjacent volcanosedimentary sequences of the Early Ordovician Snooks Arm Group (Bedard *et al*., 1996), the Sops Head Complex and the adjacent mid-late Ordovician sequences of the Wild Bight and Badger groups (O'Brien and MacDonald, 1996; O'Brien, 2000), the mid Ordovician volcano-sedimentary sequences of the western Wild Bight Group (O'Brien, 1997), and the early Ordovician South Lake Igneous Complex and adjacent Early and mid Ordovician sequences of the eastern Wild Bight Group (Swinden *et al*., 1990; MacLachlan, 1998; 1999).

TECTONIC SETTING OF THE DUNNAGE ZONE

In the Canadian Appalachians, vestiges of ancient oceanic terranes (Fyffe and Swinden, 1991) have been traditionally assigned to the lower Paleozoic Dunnage tectonostratigraphic zone (Williams, 1979) on the assumption that they originally developed in the Iapetus Ocean between the Precambrian sialic crust of the Laurentian and Gondwanan cratons (Dalziel, 1992). Oceanic crust was produced during the initial northward drift of Laurentia (McCausland and Hodych, 1998) from late Precambrian Gondwana and the subsequent detachment of other peri-Gondwanan crustal blocks (O'Brien *et al*., 1996; van Staal *et al*., 1996) from this south polar supercontinent. Some Dunnage Zone rocks may comprise suspect terranes accreted to intraoceanic tracts or to microcontinental blocks by spreading-ridge push or margin-parallel displacements; others probably originated in similar biogeographic and paleogeographic regions of the Iapetus Ocean (Williams and Hatcher, 1982; Neuman, 1984). Allochthonous and neoautochthonous rocks in the Dunnage Zone (Quinlan *et al*., 1992) formed at different times in various parts of Iapetus (Torsvik and Trench 1991; Liss *et al*., 1993; van der Pluijm *et al*., 1995).

Workers on the Laurentian and Gondwanan margins of the Dunnage Zone have postulated that the Iapetan crust and mantle preserved in the Canadian Appalachians was mostly formed by rifting and spreading within small ocean basins. Marginal basin oceanic crust was generated near continental margin or intraoceanic subduction zones at the same time that older simatic crust was being recycled in the downgoing slab (e.g., van Staal *et al*., 1991; Cawood and Suhr, 1992). This supra-subduction zone magmatism is known to have occurred on both margins of Iapetus during discrete accretionary events (Table 2), some of which led to complete or partial obduction of the various arc ophiolites (Williams and Piasecki, 1990; Suhr and Cawood, 1993; Jenner and Swinden, 1993; Bedard *et al*., 1998). The dated arc-related rocks of the Newfoundland Dunnage Zone (Dunning *et al*., 1991) are both older and younger than the mainly Early Ordovician ophiolite fragments (mostly Tremadoc ages reported by Spray and Dunning, 1991); whereas, ophiolites of the New Brunswick Dunnage Zone have ages ranging from Early Cambrian to late Middle Ordovician (van Staal *et al*., 1996) and span a longer time than the dated volcanic arc sequences (Dostal, 1989; McLeod *et al*., 1992, 1994).

Paleontological and paleomagnetic studies of certain lower Paleozoic oceanic and island arc rocks in the eastern-

[5] *Note:* Kerr (1996) advanced an alternative notion of a detached northwest-directed anticlinal nappe (with an imbricated lower fold limb) and interpreted the mainly inverted younger "terranes" of the Roberts Arm Group to have occurred at progressively lower levels in a restored pre-Silurian thrust stack.

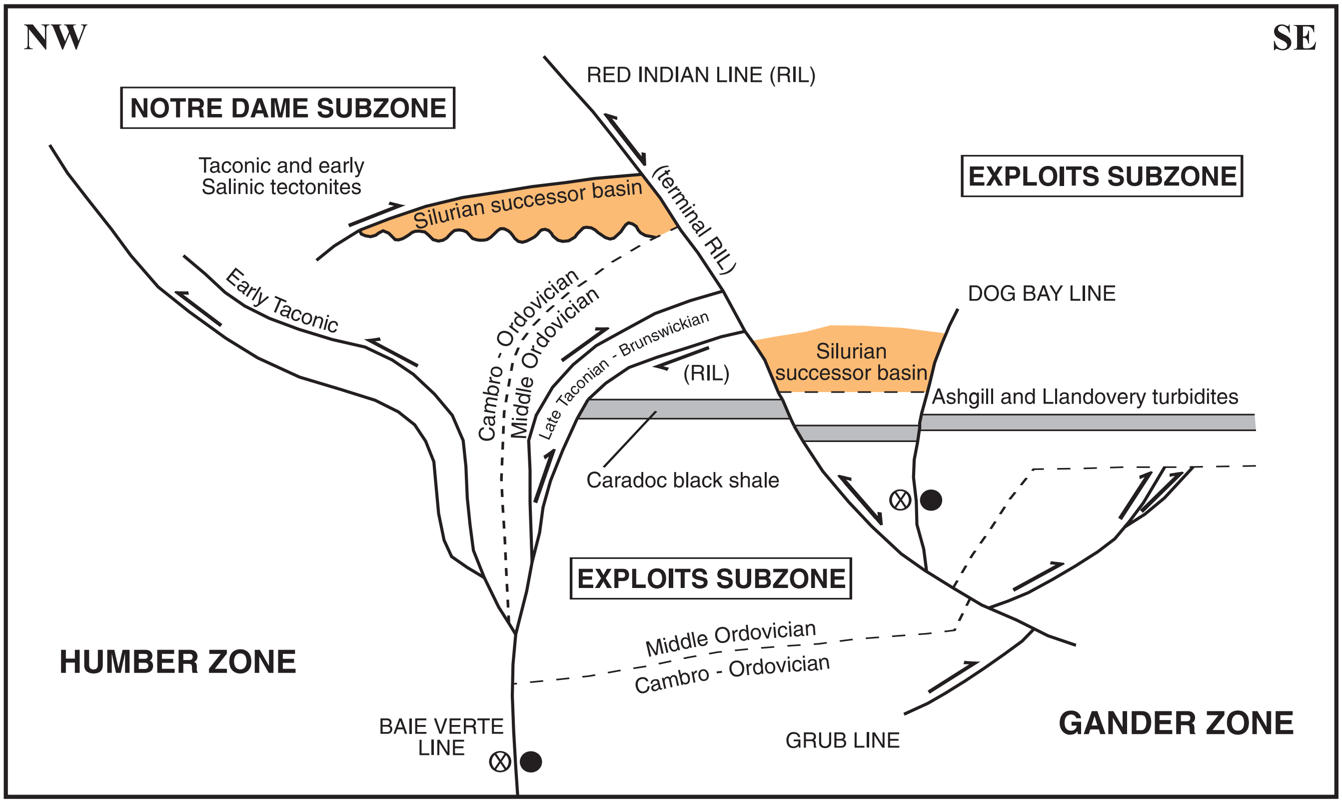


Figure 5. *Figurative cross section encapsulating an interpretation of the orogenic accretion history of the Notre Dame and Exploits subzones of the Dunnage Zone.*

most part of the Notre Dame Subzone indicate a peri-Laurentian affinity (Nowlan and Thurlow 1984; O'Brien and Szybinski, 1989; van der Pluijm *et al*., 1993). Tectonically juxtaposed are Middle Ordovician Exploits Subzone rocks thought to have been deposited in the middle of the Iapetus Ocean and Cambrian to Early Ordovician Exploits Subzone rocks deemed to have originated in the southern Iapetus Ocean (Neuman, 1988; Pickering *et al*., 1988). The Red Indian Line is probably a mid-late Ordovician boundary separating Middle Ordovician rocks of peri-Laurentian origin from those formed in mid-Iapetan or more southerly paleolatitudes (Table 2).

The Notre Dame Subzone of the Dunnage Zone interacted with the Laurentian Humber Zone in late Early Ordovician (mostly foreland) and early Middle Ordovician (mostly hinterland) phases of the Taconic Orogeny (Cawood *et al*., 1995; Table 2). Some Notre Dame Subzone rocks document evidence of geochemical interaction with Laurentian continental crust as early as the Cambrian (Szybinski 1995) or Early Ordovician (Dec *et al*., 1997), long before the obduction of Dunnage ophiolites onto the drowned Humber continental margin (H. Williams and Cawood, 1989; Knight *et al*., 1991). The fact that the Red Indian Line approximately overlies what is probably the limit of sub-Dunnage Zone Grenvillian basement at mid-crustal depths (Keen *et al*., 1986), and that probable peri-Gondwanan Central Block basement occurs immediately southeast of the line (Marillier *et al*., 1989), implies a movement history on the Red Indian Line which continued until terminal continent–continent collision (Table 2; Figure 5).

In Newfoundland, the widest age range of dated Dunnage Zone rocks is found on the peri-Gondwanan margin of Iapetus within the thick supracrustal deposits of the Exploits Subzone (Evans *et al*., 1992). There, the boundary between lower and mid Paleozoic deep-marine strata has been long known to be regionally conformable (*see* Previous Geological Work; Figure 5). Disjunct-arc and ophiolite complexes (Colman-Sadd *et al*., 1992a and references therein) originated in Iapetus close to the thinned continental margin of Gondwana (MacLachlan and Dunning, 1998a), and so contained faunas typical of the southern continent and its borderlands (S.H. Williams *et al*., 1992a).

These Cambro-Ordovician intraoceanic rocks were conformably and, in places unconformably, overlain by late Early and early Middle Ordovician arc, arc rift, back arc and melange complexes (e.g., Hibbard and H. Williams, 1979; Evans *et al*., 1990; van Staal, 1994; S.H. Williams and Tallman, 1995; Lee and H. Williams, 1995; O'Brien *et al*., 1997; MacLachlan and Dunning, 1998b). The Early–Middle Ordovician rocks formed near or above the peri-Gondwanan margin of Iapetus, which was partly oceanic and partly continental at that time. All these Exploits Subzone rocks were succeeded by late Middle Ordovician fossiliferous sedimentary strata (and minor volcanic rocks) that yield, for the first time, a North American faunal assemblage (S.H. Williams *et al*., 1995). The Middle Ordovician successions in the upper parts of the Wild Bight, Exploits, Summerford and Victoria Lake groups (Elliot *et al*., 1989; MacLachlan *et al*., 1998; Figure 42) and farther east within the upper parts of the Baie d'Espoir and Davidsville groups (Colman-Sadd, 1980; Currie, 1995c) thicken and coarsen northwestward above various components of the northern Exploits Subzone.

Volcanic and hypabyssal rocks of post-Arenig, pre-Caradoc age are most abundant near the Red Indian Line and decrease in proportion to intraformational sedimentary rocks southeastward approaching the Dunnage–Gander composite margin. An exception may possibly have been along the peri-Gondwanan basement promontory of southwest Newfoundland (Schofield *et al*., 1998). To the south and possibly at greater depth in the east, the Ganderian and Avalonian substrate of the Exploits Subzone was still being affected by syntectonic plutonism in the late Middle Ordovician (Colman-Sadd *et al*., 1992b; O'Brien *et al*., 1993; Table 2), despite being in a distal position relative to the active volcanic arc front.

The oldest deep-marine deposits lying above the Middle Ordovician and older Exploits Subzone rocks (Williams, 1995b) occur in relatively contiguous sedimentary successions devoid of interstratified volcanic rocks. Ranging from early Late Ordovician to late Early Silurian in age, they represent the synorogenic infill of remnant seaways in the constricted Iapetus Ocean (Williams and O'Brien, 1991). In various places, a relatively thick graptolite-bearing siliciclastic flysch or a relatively thin, conodont-bearing calcareous flysch lie both conformably and disconformably above the Middle Ordovician and older tectonic components of the Exploits Subzone (*see* Table 1). Based on their detrital composition (Nelson and Casey, 1979) and paleocurrent analysis (Arnott *et al*., 1985; Pickering, 1987), these turbidites were interpreted to have accumulated in dynamic basins adjacent to the uplifted peri-Laurentian magmatic arc (Colman-Sadd *et al*., 1992b; H. Williams, 1995c). Dec *et al*. (1993) interpreted some detritus in these deep turbidite troughs to have been recycled from shallow-marine clastic formations and partially lithified carbonate reefs which had originally covered certain Middle Ordovician rocks presently assigned to the southeasternmost Notre Dame Subzone and the northwesternmost Exploits Subzone. In the Late Ordovician deposits covering the Exploits Subzone, the endemic Atlantic Region conodont faunas of the southern Iapetan borderlands (specifically Baltica) were mixed with local influxes of Midcontinent Laurentian faunas (Nowlan *et al*., 1997), presumably as Laurentia had already collided with Baltica farther northeast in the Appalachian–Caledonian orogenic belt (Murphy *et al*., 1996).

The Silurian terrestrial deposits that cover parts of the Dunnage Zone have been generally interpreted as the products of erosion that followed Taconian mountain building, as the Penobscot mountains were already bevelled and presumably undersea in the Silurian (Table 2; Figure 5). Thus, workers have traditionally contrasted the terrigenous cover of the Notre Dame Subzone, which unconformably overlies variably deformed basement units of Middle Ordovician and older marine rocks, with their paraconformable terrestrial and marine correlatives in the Exploits Subzone (Figures 3 and 5; Table 1). In Notre Dame Bay, mid-Paleozoic deposition of redbeds and terrestrial volcanics has been postulated to have occurred after postorogenic emergence and isostatic readjustment of the Notre Dame Subzone in the Late Ordovician (cause of non-deposition of the Badger Group) or subsequent to an Early Silurian period of thrust-related uplift in the Exploits Subzone (cause of facies variations and melange development in the Badger Group).

Whether it is above unconformity bounded Silurian sequences or within paraconformable Ordovician–Silurian sequences, primary depositional boundaries between Silurian terrestrial and marine strata are preserved inside several stratigraphic units. Such boundaries are observed in Silurian successions lying above the Notre Dame and Exploits subzones of the Dunnage Zone. However, terrestrial and marine rocks developed at different stages of the Silurian in different areas. Local non-sequences within terrestrial and shallow-marine Silurian successions (Sops Arm, White Bay) and local hiatuses in deep-sea deposition near the Ordovician–Silurian boundary (Goshen, New World Island) are generally too young to represent the Taconic unconformity and are probably more closely related to changes in Silurian sea level or Silurian tectonic regime.

The Mid Paleozoic development of the orogen (Table 2) is commonly viewed as the history of isolated successor basins formed above different ages of variably remobilized Laurentian and Gondwanan basement (H. Williams, 1995d). Basin distribution is unrelated to the Early Paleozoic zonation of the orogen (H. Williams, 1995a,b); however, accumulation sites were controlled, in places, by reactivated Ordovician and older structures (e.g., O'Brien *et al*., 1993).

Accretionary History of the Exploits Subzone

Rocks of the Exploits Subzone preserve a complex and protracted record of orogenic accretion and tectonic assembly (Table 2). In the Early Ordovician (mid-late Arenig), at least some of the oceanic rocks in this subzone were emplaced upon the Gander Zone continental margin in the Penobscot event (Williams and Piasecki, 1990; Colman-Sadd *et al*., 1992b), although the extent to which this collision affected the peri-Gondwanan Iapetan deposits and the Avalonian basement inliers is not fully understood (O'Brien *et al.*, 1993; Tucker *et al.*, 1994; O'Brien *et al.*, 1997). In the northernmost Exploits Subzone, a soft collision in the Early Silurian produced submarine uplift and associated melange tracts (Reusch, 1987) that are generally regarded as marking the initiation of closure of the Exploits arc–back arc complex (Lafrance and P. Williams, 1992; Currie and H. Williams, 1993; Currie, 1995a).

The abundant mafic dyke swarms of the Exploits Subzone and most of its sill complexes were intruded episodically during the Tremadoc, Arenig , Llanvirn and Llandeilo. They have been generally interpreted to reflect arc extension, possibly in association with subduction zone steepening or polarity reversals during these subdivisions of the Ordovician (O'Brien *et al*., 1997; MacLachlan and Dunning, 1998b). As an early Middle Ordovician island arc evolved along the western margin of the Exploits Subzone, the late Arenig and younger intrusive bodies in this suite were preferentially emplaced into the emergent regions of the active volcanic arc, the flanks of a remnant arc and a volcanosedimentary rift basin within this arc complex (MacLachlan, 1998). Some may have been intruded when this part of the Exploits Subzone tectonically encroached upon age-equivalent, magmatic arc rocks of the eastern Notre Dame Subzone (van Staal, 1994; Cawood *et al*., 1995; Whalen *et al*., 1996).

Although volcanic, sedimentary, plutonic and metamorphic rocks of Early and Late Silurian age are most widespread in central Newfoundland (Table 1), tectonism related to mantle subduction has been generally discounted in the Dunnage Zone at this time (Colman-Sadd *et al*., 1992a; Cawood *et al*., 1995). Ashgill and Llandovery magmatic arc rocks are locally emplaced into the easternmost Notre Dame Subzone (Rainy Lake Complex of Whalen, 1989) and the westernmost Exploits Subzone [Older Granites of the Mount Peyton (Dickson, 1992) and Hodges Hill (Dickson, 2000) batholiths]. Like their correlatives in the Bronson Hill arc of New England (Tucker and Robinson, 1990), these arc-related rocks may have been generated above long-lived subduction zones in Iapetan reentrants or, alternatively, they may reflect Mid Paleozoic melting of fossilized Early Paleozoic mantle (Whalen *et al*., 1994).

Some relict Iapetan depocentres may have been influenced in the mid-Paleozoic by far-field effects of the active margin of the Rheic Ocean, which lay to the south of Avalonia (Keppie *et al*., 1991; Keppie, 1993; Table 2). In Newfoundland, Nova Scotia and New Brunswick, Late Ordovician and Early Silurian volcanic and plutonic rocks on the Avalonian margin of the Appalachian Orogen have been commonly postulated to have originated during active continental-margin subduction (Greenough *et al*., 1993; Johnson and McLeod, 1996; Bergstrom *et al*., 1997).

In central Newfoundland, terrestrial strata were deposited and deformed in the Silurian (Dunning *et al*., 1990), when the composite terranes of Gondwana continued to obliquely converge upon the accreted terranes of mid Ordovician Laurentia. Individual depocentres record the effects of volcanism or plutonism during discrete transtensive and transpressive stages of basin development between the late Llandovery and the late Pridoli (e.g., Coyle and Strong, 1987). The Silurian evolution of terrestrial strata and cogenetic subvolcanic intrusions was quite varied and strongly influenced by the local effects of the underlying Silurian infrastructure (O'Brien, 1998b). In the Late Silurian, collision of Laurentian and Gondwanan continental basement blocks produced the climactic orogeny of the Newfoundland Appalachians (Stockmal *et al*., 1990), which affected terrestrial and marine strata throughout the high- and low-grade parts of the Dunnage Zone (Elliot *et al*., 1991; Cawood *et al*., 1994).

PRESENT INVESTIGATION

Previous systematic surveys of central Notre Dame Bay have identified particular rock units which have been mineralized in unique environments. Some rock units have been altered in more than one paleotectonic regime during separate geological events (e.g., syngenetic Ordovician base-metal mineralization followed by epigenetic Silurian precious-metal mineralization). Tracking the history of such prospective rocks has been hampered by remaining gaps in our knowledge of the regional geology of the type area of the northern Dunnage Zone. The main objective of this study is to provide detailed geological information about selected well-exposed areas in Notre Dame Bay with an aim to resolving some outstanding problems concerning the origin and evolution of the region's Ordovician and Silurian rocks. Parts of several 1:50 000 scale geological maps were produced based on this information (O'Brien, 1990, 1991a, b, 1992b, 1993a).

In addition to detailed mapping, studies have been carried out to investigate the stratigraphy, structure, sedimentology, lithogeochemistry, paleontology and geochronology of the local Paleozoic sequences found in this part of the Dunnage Zone. Most of this geological information is thought to be applicable to correlative units in the poorly-exposed interior of the island, where many of these units are observed to be altered or mineralized. In the Exploits Subzone, emphasis has been placed on the lithostratigraphic subdivision of the Exploits Group and the relationships of its constituent formations and members to local subunits of the southwestern Dunnage Melange, the easternmost Wild Bight Group, and the formations of the type Botwood Group. The regional biostratigraphy and structure of the Point Leamington Formation has been examined and compared with other Badger Group units farther east. Comparisons and distinctions have also been made amongst the Lawrence Harbour and Shoal Arm formations and several other informal units having a substantial component of graptolitic black shale. Mafic and felsic plutonic rocks of various ages have been mapped; pretectonic varieties have been emphasized and those intruded or tectonically emplaced into Ordovician and Silurian host rocks are described in some detail.

In the Notre Dame Subzone, the constituent formations of the Cottrells Cove Group have been mapped and logged, and an internal lithostratigraphy has been proposed on the basis of related chemostratigraphic and sediment provenance studies. Various structural and stratigraphic relationships between the Cottrells Cove Group and the uppermost

Western Head Formation of the Moretons Harbour Group are described. The Red Indian Line structural zone has been mapped on the Fortune Harbour Peninsula and in the Bay of Exploits, and its regional geometry has been portrayed in cross sections. These illustrate the relative setting of the block-in-matrix rocks of the Boones Point Complex and the Dunnage Melange, depict their individual relations with unbroken Middle and Upper Ordovician map units and generally show the features of regional deformation that they share with early Silurian and older strata in central Notre Dame Bay.

LITHOSTRATIGRAPHY OF THE NOTRE DAME AND EXPLOITS SUBZONES IN CENTRAL NOTRE DAME BAY

The New Bay–Bay of Exploits area in north-central Newfoundland is underlain, for the most part, by lower Paleozoic strata belonging to the Exploits and Notre Dame subzones of the Dunnage Zone (Williams, H. *et al*., 1988; Swinden *et al*., 1988; Williams, H., 1995b). All these rocks are deformed by a regional Z-shaped oroclinal fold termed the Notre Dame Bay flexure (Figure 3). The Red Indian Line, which trends northeastward and northwestward on the limbs of this oroclinal flexure, is manifested within the map area as a 2 to 3 km wide mylonite zone that contains discontinuous belts of olistostromal and tectonic melange. Most observations indicate that the regional stratigraphic facing direction of Exploits Subzone units is westward or northward; whereas, the regional stratigraphic facing direction of Notre Dame Subzone units is eastward or southward.

GENERAL STATEMENT

In the area surveyed, Cambro-Ordovician volcanic and minor sedimentary rocks comprise the stratified units of the Notre Dame Subzone. These mainly unfossiliferous strata are assigned to the Sweeney Island and Western Head formations of the Moretons Harbour Group and the Fortune Harbour and Moores Cove formations of the Cottrells Cove Group (Figures 3 and 4; Table 1). The Boones Point Complex is most closely associated with the Moores Cove Formation and has been previously included in the Cottrells Cove Group.

In central Notre Dame Bay, the fossil-bearing lithostratigraphic units of the Exploits Subzone are composed of Middle Ordovician and older sedimentary and volcanic rocks. Most belong to the Exploits Group but some reside in the easternmost Wild Bight Group (Figure 3; Table 1). From oldest to youngest, the Exploits Group is made up of the Tea Arm, Saunders Cove, New Bay and Lawrence Head formations and the overlying Strong Island chert and Hummuck Island limestone units. In the map area, the Wild Bight Group is mainly represented by the youngest internal unit, the Pennys Brook Formation.

A regional overstep sequence of the Middle Ordovician to Late Silurian sedimentary and volcanic rocks lies above some of the stratified units of the Exploits Subzone (Table 1). Middle to Late Ordovician sedimentary strata occur in the Shoal Arm, Lawrence Harbour and Luscombe formations. The succeeding units of the Late Ordovician–Early Silurian Badger Group include the Point Leamington greywacke, the Goldson conglomerate, the Randels Cove conglomerate, the Campbellton greywacke, the Lewisporte conglomerate and the Upper Black Island greywacke (Figure 3). The youngest constituents of the overstep sequence are the terrestrial Lawrenceton and Wigwam formations of the Early–Late Silurian Botwood Group.

NOTRE DAME SUBZONE

Moretons Harbour Group

The Ordovician or older volcanic rocks of the Moretons Harbour Group (Table 1) occur north of the Chanceport Fault within the study area and throughout the Notre Dame Bay region (Figure 3).

Distribution and Thickness

In the area surveyed, the Moretons Harbour Group is restricted to the northern tip of the Fortune Harbour peninsula and is well exposed on the coastal headlands (Figure 3). There, the minimum stratigraphic thickness is about 1.5 km.

Stratigraphic Nomenclature

The Moretons Harbour Group consists of two lithologically distinct formations on the Fortune Harbour peninsula. These are the older Sweeny Island Formation and the younger Western Head Formation (Dean, 1977; Figure 6). Prior to Dean's mapping and stratigraphic revision, such strata were assigned to the northern tract of Helwig's (1967) Lushs Bight Group.

Lithostratigraphy

Neither the stratigraphic base nor the stratigraphic top of the Moretons Harbour Group is exposed in the area surveyed. At several coastal localities in the Webber Bight–Western Head Cove area of the Fortune Harbour peninsula, the openly folded contact of the Sweeny Island and Western Head formations is observable and there the boundary is stratigraphic and conformable (Figure 6). The Sweeny Island Formation everywhere underlies the Western Head Formation and, where fully displayed, the gradational transition zone is up to 15 m thick.

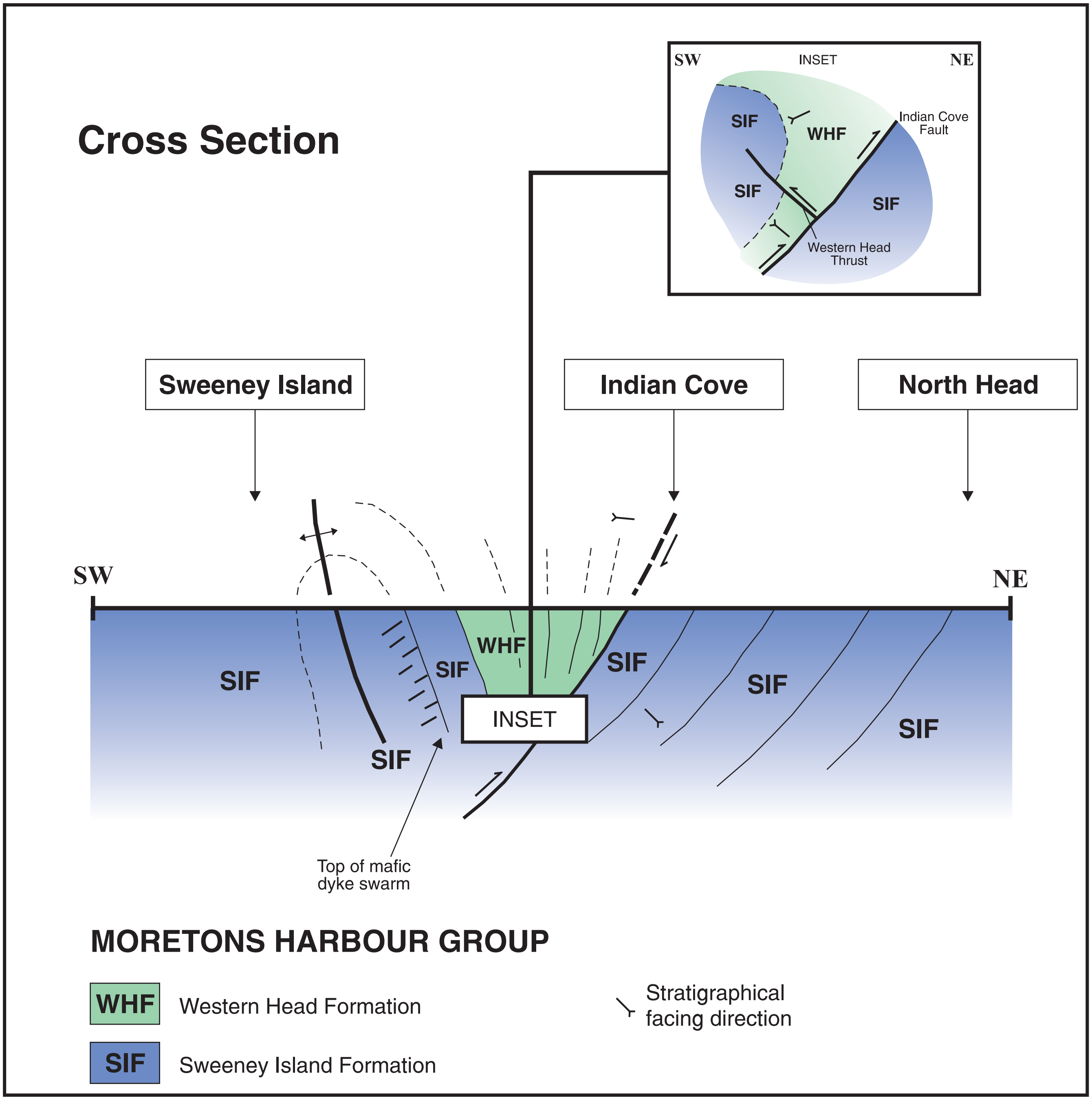


Figure 6. *Simplified northeast–southwest cross section of the Sweeney Island and Western Head formations of the Moretons Harbour Group between North Head and Sweeney Island, Fortune Harbour. Inset illustrates the displacement of a previously overfolded succession of formations along the thrust observed at the eastern headland of Western Head Cove [E627250 N5488750] and the inferred geometrical relationship of this thrust to the underlying Indian Cove Fault.*

A regional anticline, which controls the disposition of the formations of the Moretons Harbour Group between the Indian Cove and Chanceport faults, plunges gently southeast (Figures 3 and 6). However, as a generalization, units of the Sweeny Island and Western Head formations are moderately dipping in most parts of the map area. The distribution of the upper subunit of the Western Head Formation indicates that the youngest parts of the Moretons Harbour Group are found in the extreme southeast of that formation's outcrop area (O'Brien, 1990). Therefore, in places, the highest preserved part of the Moretons Harbour Group lies adjacent to the Cottrells Cove Group (Dec *et al*., 1997).

Lithology

The Sweeny Island Formation is composed of coarse grained, porphyritic, basaltic lavas which host ubiquitous diabase dykes that, in certain locations, are intensely sheeted. Where numerous, these vertical intrusions separate very narrow screens of gently dipping country rocks. In some places, the Sweeny Island Formation contains very thick, poorly layered horizons comprising individual flow units; in other localities, it lacks internal volcanic stratification. Pillow lava and inter-lava sediment are notably absent in the Sweeny Island Formation. However, such rocks are typical of several correlative or younger formations of the Moretons Harbour Group (Table 1), which outcrop on Notre Dame Bay islands farther to the northeast.

In ascending order, the Western Head Formation of the Moretons Harbour Group is composed of a lower subunit of mafic agglomerate and pillow breccia, an intermediate subunit of spectacular mafic pillow lava, and an upper subunit of light grey siliceous argillite, minor grey and red chert, and rare green-grey epiclastic wacke (O'Brien, 1990). The younger formation of the Moretons Harbour Group appears not to contain the abundant swarms of sheeted hypabyssal rocks characteristic of the older formation. The Western Head Formation is, however, host to a variety of multiple or composite minor intrusions and some of these, particularly near the base of the unit, may belong to the upper part of the Sweeny Island dyke swarm.

Age

In the area surveyed, all external boundaries of the Moretons Harbour Group are faulted. Thus, the original stratigraphic relations of the Moretons Harbour Group to other rock groups are concealed and its absolute age is unknown. Its regional relationships to the dated Late Cambrian, Early Ordovician and Mid Ordovician plutonic and volcanic rock units in the Notre Dame Bay sector of the Notre Dame Subzone are equivocal, though Swinden *et al*. (1998) suggested a correlation with the Taconian-deformed Lushs Bight Group.

Regional Interpretations

Dean (1978) interpreted the Sweeny Island Formation as representing a part of the oceanic crustal sequence of a Lower Ordovician ophiolite (then thought to be Dunnage Zone basement) and deemed it to be the oldest rock unit exposed on the Fortune Harbour peninsula (in what is now the Notre Dame Subzone). The gross lithology and lithodemic associations of the subunits of the Western Head Formation of the Moretons Harbour Group are, nevertheless, strikingly similar to those of the Fortune Harbour Formation of the Cottrells Cove Group. The Fortune Harbour Formation comprises the southerly adjacent tract of Notre Dame Subzone rocks (Figure 3) and includes felsic and mafic volcanic rocks of late Tremadoc–early Arenig age (Dec *et al*., 1997). It is possible that the Western Head Formation could be the partial or whole equivalent of the Fortune Harbour Formation, although the association of pillow lava, chert and volcanogenic sediment is a feature common to most of the Middle Ordovician and older rock units in Notre Dame Bay. However, if this correlation is correct, then an inconsequential amount of displacement has occurred on the fault at the southern boundary of the Moretons Harbour Group, i.e., the Chanceport Fault of Dean and Strong (1977). In contrast, on paleomagnetic considerations (Johnson *et al*., 1991), the Moretons Harbour Group originated in a completely different part of the Ordovician Iapetus Ocean than the Cottrells Cove Group. Thus, in the latter interpretation, the Chanceport Fault is equal or greater in status to the fault at the southern boundary of the Cottrells Cove Group (the Lukes Arm Fault in Figure 3).

An important ramification of the partial correlation of the Moretons Harbour and Cottrells Cove groups would be that the rocks of the Notre Dame Subzone on the Fortune Harbour peninsula young regionally southward, although internal fault-bounded panels within the subzone contain strata that are commonly inverted and northward-younging (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65). In this tectonic scenario, the regional stratigraphic facing confrontation in the Dunnage Zone would then be at the Lukes Arm Fault, because Exploits Subzone rocks face regionally northward south of that fault.

Cottrells Cove Group

The main tract of the Lower–Middle Ordovician Cottrells Cove Group displays well-preserved syndepositional features and occurs north of the Lukes Arm Fault (the Red Indian Line in Figure 3). However, recognizable tectonic fragments of this rock group occur farther south within the Red Indian Line structural zone in association with discontinuous belts of mylonite and block-in-matrix melange (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65).

Distribution and Thickness

The Cottrells Cove Group forms a 16 km long, northwest-trending belt between Fleury Bight and North Harbour on the Fortune Harbour peninsula. It also comprises a 7 km long, northeast-trending belt on the nearby Exploits, Grassy, Duck, Swan and Hornet islands in the western Bay of Exploits. The total stratigraphic and structural thickness of the Cottrells Cove Group is in the order of 6 km combined.

Stratigraphic Nomenclature

As described in this report, the south-facing Cottrells Cove Group contains two formations, the Fortune Harbour Formation and the Moores Cove Formation. These units have been both successfully subdivided in well-exposed coastal sections and detailed stratigraphic logs of the distinctive volcanic and sedimentary rocks in these formations have been formulated (Dec and Swinden, 1994).

Lithostratigraphy

Where exposed, the mutual contact of the Fortune Harbour and Moores Cove formations is everywhere faulted. However, based on sediment provenance and age determinations (Dec *et al*., 1997), the older Fortune Harbour Formation is thought to have stratigraphically underlain the Moores Cove Formation. As mapped and defined herein, neither the stratigraphic base nor the stratigraphic top of the Cottrells Cove Group is exposed.

Stratigraphic Revision

As originally defined by Dean (1977), the Cottrells Cove Group comprised three conformable formations. From south to north, and from oldest to youngest, these were the olistostrome-dominant Boones Point Complex, the turbidite-dominant Moores Cove Formation and the volcanic-dominant Fortune Harbour Formation. In this report, the writer has excluded the Boones Point Complex from the Cottrells Cove Group and has reversed the stratigraphic order of the Moores Cove and Fortune Harbour formations.

In the past, the Cottrells Cove Group was held to have been originally stratigraphically continuous with an underlying olistostrome-bearing, wacke-dominated succession (Helwig's Point Leamington Greywacke). Because this succession was previously known to be Late Ordovician in age, the north-facing Cottrells Cove Group was assigned a probable Early Silurian age by Dean (1977) and Kean *et al*. (1981). In this report, the Cottrells Cove Group is deemed to be older than the Point Leamington Formation of the Badger Group and the boundary between these units is thought to be a Silurian or younger fault (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65).

Lithology

The mafic volcanic-dominated Fortune Harbour Formation is made up of basaltic agglomerate and pillow breccia, numerous spectacular pillow lava units, and fine grained felsic tuff and rhyolitic agglomerate. The felsic volcanic rocks are interbedded with laminated siliceous argillite and conspicuous red and grey chert horizons (O'Brien, 1990; Plates 1, 2, 3 and 4).

The turbidite-dominated Moores Cove Formation is composed of distinctly weathered, feldspathic and lithic wacke interbedded with variable amounts of grey argillite and rarer chert (Plates 5 and 6). The younger formation has a very thin basal unit of pillow lava with intercalated fossil-bearing limestone which lies immediately below the ferruginous and calcareous wackes. The Moores Cove Formation is devoid of the thick sequence of mafic volcanic rocks and interstratified turbidite lenticles that characterize the Fortune Harbour Formation.

Plate 1. *Flow-layering wrapping around a resistant block of mafic–felsic volcanic breccia in a crystal-rich ignimbrite in the Fortune Harbour Formation of the Cottrells Cove Group. Such coarse-grained pyroclastic flows are restricted to a felsic volcanic centre within the formation located near Fleury Bight.*

Viewed regionally, the epiclastic turbidites of the Moores Cove Formation are thin-bedded, fine-grained and most siliceous near the stratigraphic base of the formation and become thicker bedded, coarser grained and more proximal in aspect towards the top of the exposed succession.

Age

Conodonts from thin Moores Cove limestones on Tinker Island, situated above pillowed basalt and below thin-bedded siliceous wacke, are latest Arenig or early Llanvirn in age (G. Nowlan, personal communication, 1998). Rounded boulders of limestone from polymictic conglomerate, which is found near the exposed top of the Moores Cove Formation on the Duck Islands (Dec and Swinden, 1994), yielded conodonts of early Llanvirn age (G. Nowlan, per-

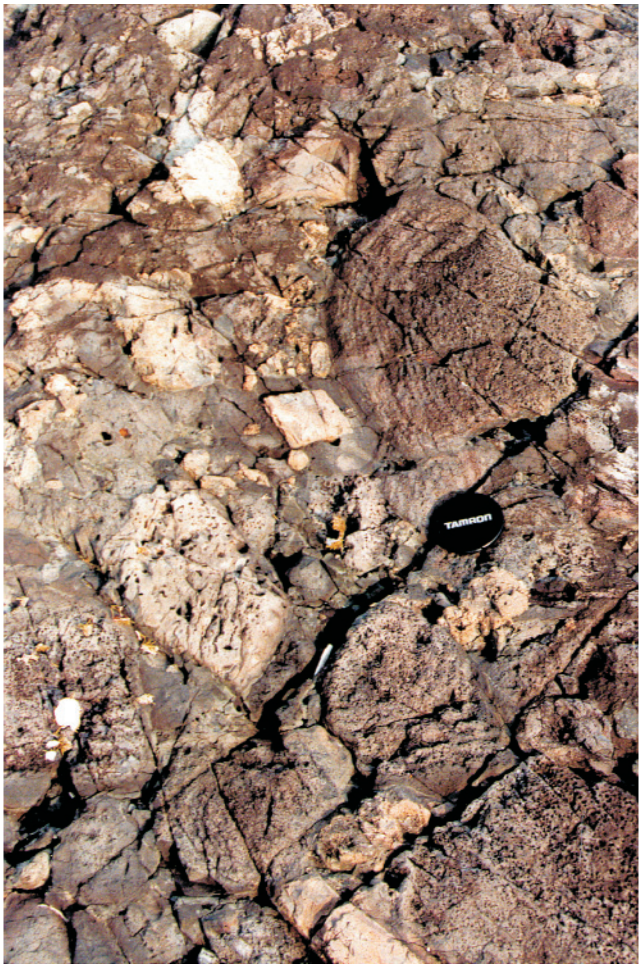


Plate 2. *A broken piece of the chilled margin of a pillow lava, coarse blocks of vesicular basalt, and angular fragments of buff felsic tuff occur in a breccia horizon near Fleury Bight, indicating coeval mafic and felsic magmatism in the Fortune Harbour Formation*

sonal communication, 1999). The fossil ages on the Moores Cove Formation, when compared with the early Arenig U/Pb age of felsic tuffs in the Fortune Harbour Formation (Dec *et al*., 1997), independently establish the depositional order of the constituent formations of the Cottrells Cove Group.

The Caradoc black shales and Ashgill sandstone turbidites, which are present in the lower part of the Exploit Subzone overstep sequence to the immediate south, are distinctly absent in the Cottrells Cove Group. Nevertheless, limestone-covered Middle Ordovician basalts of mid ocean ridge affinity form the depositional substrate of both the Moores Cove succession and the Lawrence Harbour–Point Leamington succession.

Correlation

The stratigraphic development and lithologic character of the internal units of the Moores Cove Formation of the Cottrells Cove Group are very similar to that of the sandstone turbidites of the Sansom Formation and the conglomeratic turbidites of Goldson Formation in the Badger Group (Table 1). In the past, this has lead to confusion in separating and mapping the discrete turbidite belts that are present along the southern margin of the Notre Dame Subzone and the northern margin of the Exploits Subzone. In particular, it has also enforced the historical view that the volcanic- and turbidite-dominated units of the Notre Dame Subzone on the Fortune Harbour peninsula faced northward and were Late Ordovician or Early Silurian in age. The writer maintains that the upper Cottrells Cove Group–lower Badger Group correlation is incorrect and that the notion of stratigraphic pinch-out of the graptolitic black shale facies to the north is flawed. Furthermore, the tholeiitic basalts, conodont-bearing limestones, epiclastic wackes and polymict boulder conglomerates of the early Middle Ordovician Moores Cove Formation of the Cottrells Cove Group are probably best correlated with similar rock types in the Crescent Lake Formation of the Roberts Arm Group (*see* Introduction, page 1).

Regional Interpretations

The oldest mafic volcanic rocks, the felsic volcanic centre and the associated cupriferous stockworks occur chiefly in the north (O'Brien, 1990; 1991b), where the volcanic and sedimentary strata of the Fortune Harbour Formation are most extensive (Figure 7). By comparison, the thickest preserved sections of epiclastic sedimentary rocks are developed within the southern outcrop area of the Moores Cove Formation (Figures 7 and 8). Both formations are, however, locally present along the northern and southern margins of the Cottrells Cove Group.

Regionally, the Cottrells Cove Group faces southward toward the Lukes Arm Fault but the group may face northward over a much smaller tract of ground near the Chanceport Fault (Figure 9). Thus, the Fortune Harbour Formation probably outcrops in a regional anticlinorium flanked by faulted synclinoria that control the disposition of the Moores Cove Formation.

In the north of the map area, near the Grassy Islands and on the northeast coast of the Fortune Harbour peninsula, a relatively thin Moores Cove succession was deposited above a relatively thick sediment-dominated succession of the Fortune Harbour Formation (Figure 9; Dec and Swinden, 1994). A more substantial thickness of the middle and upper Moores Cove Formation is preserved on the southern margin of the Cottrells Cove Group than is observed along its northern boundary (Figure 7; O'Brien, 1991b), even though the basal Moores Cove Formation is exposed in both regions.

In the south of the map area, the greyish-green turbidites of the Fortune Harbour Formation are conspicuously absent in several locations. For example, in the Rowsells Cove area (O'Brien *et al*., 1994; Figure 7), the basal basalt of the Moores Cove Formation is stratigraphically succeeded by green siliceous wacke and red laminated argillite near

Plate 3. *Near the village of Fortune Harbour, thin-bedded, buff felsic tuffs (right side of photograph) are interbedded with laminated red cherts (centre of photograph) and medium-bedded, light-green wackes (left side of photograph). Typical of the middle–upper part of the Fortune Harbour Formation, similar felsic tuffs have an Early Arenig absolute age and overlie calc-alkaline pillow lavas.*

Plate 4. *Fine-grained, quartz-feldspar crystal tuffs of the Fortune Harbour Formation contain scattered rhyolite fragments surrounded by distinct haloes in the adjacent matrix, possibly indicating that they were hotly erupted bombs or that they were reactive with percolating matrix fluids.*

the faulted boundary with the Fortune Harbour Formation (O'Brien, 1990; Figure 7). Here, a relatively thick and well-exposed sequence of Fortune Harbour calc-alkaline basalts is completely devoid of volcanic-derived turbidite lenticles.

Reassigning the Moores Cove Formation to a position stratigraphically above the Fortune Harbour Formation has important implications for the Ordovician paleogeography and Silurian structure of the Cottrells Cove Group. The map pattern described above may have been caused by the primary thickening of sediment in opposing directions in the two formations, by the secondary effect of a relatively larger amount of tectonic removal of Moores Cove strata on the northern margin of the group, or by some combination of both phenomena. These factors probably contributed to the marked regional asymmetry of the Cottrells Cove sedimentary basin.

Tectonic Implications

In several locations, right-way-up strata in the middle part of the Fortune Harbour Formation of the Cottrells Cove Group face, in the stratigraphic sense, away from the underlying Moretons Harbour Group (Figure 10). In some sections across the Chanceport Fault, however, it appears that the youngest observed strata in the Fortune Harbour Formation young toward the Moretons Harbour Group (Figure 9). If the south-facing volcanic and sedimentary rocks lying above the Sweeny Island Formation everywhere belong to the Moretons Harbour Group, then an anomalously large amount of stratigraphic separation must occur across the Chanceport Fault, since all of the Moores Cove Formation and some of the Fortune Harbour Formation would have had to have been structurally excised. Alternatively, if the right-way-up, south-facing rocks that lie north of the Chanceport Fault are more appropriately grouped with the Fortune Harbour Formation than with the Western Head Formation, then a relatively thin succession of the Cottrells Cove Group must have been originally deposited above the Sweeny Island Formation of the Moretons Harbour Group.

Boones Point Complex

The Boones Point Complex is an inhomogeneously deformed, dark grey argillite-hosted, scaly foliated, block-in-matrix melange unit that is spatially associated with the Lukes Arm fault zone (Figure 3).

Distribution and Thickness

The complex extends discontinuously some 17 km from Woody and Green islands near the community of Leading Tickles southeastward to Little Grego Island off the east coast of the Fortune Harbour peninsula. A maximum structural thickness of approximately 900 m can be estimated for the largest fault-bounded tract of the Boones Point Complex (the Southeast Arm panel in Figure 7).

Stratigraphic Nomenclature

The name Boones Point Complex was formally proposed by Dean (1978), although parts of the unit were originally mapped by Helwig (1967). Both workers considered the complex to be characterized by pebbly mudstone units distinguished by outsized blocks of stratified and intrusive rocks; however, they also included various partially broken formations of sedimentary and volcanic strata.

Plate 5. *North of Little Grego Island, laterally continuous, fine-grained, grey sandstone turbidites are interstratified with thinly bedded, red and green, siltstone turbidites and laminated red mudstones. Characteristic of the lower part of the Moores Cove Formation, such strata occur immediately above a basal section of pillowed basalts and intercalated conodont-bearing limestones.*

Stratigraphic Revision

In this report, only the block-in-matrix melange tracts have been grouped in the Boones Point Complex. The coherently-bedded strata that lie tectonically adjacent to the Boones Point Complex have been assigned to constituent units of the Cottrells Cove and Exploits groups and, less commonly, to the Point Leamington Formation.

Plate 6. *Thick-bedded, grey–green, volcaniclastic wackes of the Moores Cove Formation are interstratified with red laminated mudstone along the southwest coast of Cottrells Cove. They become increasing dominant as Moores Cove strata coarsen and thicken upward.*

The Boones Point Complex was deemed by earlier workers to have originally occupied a stratigraphic position between the Goldson Formation (now part of the overstep sequence of the Exploits Subzone) and the Moores Cove Formation (now the youngest unit of the Notre Dame Subzone on the Fortune Harbour peninsula). Accordingly, Dean (1978) and Kean *et al*. (1981) considered this map unit to be either Late Ordovician or Early Silurian, as they knew that olistostromal sequences in other parts of Notre Dame Bay spanned this same time interval. However, the complex does not represent the basal stratigraphic unit of Dean's (1978) Cottrells Cove Group (*see* Age and Correlation, page 105).

Lithology

The outcrop areas of the dark grey argillite-hosted, iron oxide-weathered melange tracts of the Boones Point Complex are kilometric in scale (Figure 7). The most obvious feature of the pebbly mudstone and block-in-matrix melange is the variable composition and size of the constituent blocks. In addition to the ubiquitous epiclastic wacke blocks, these include extrusive and intrusive magmatic rocks, hemipelagic and chemical sedimentary rocks, and deep-sea limestone associated with mafic volcanic

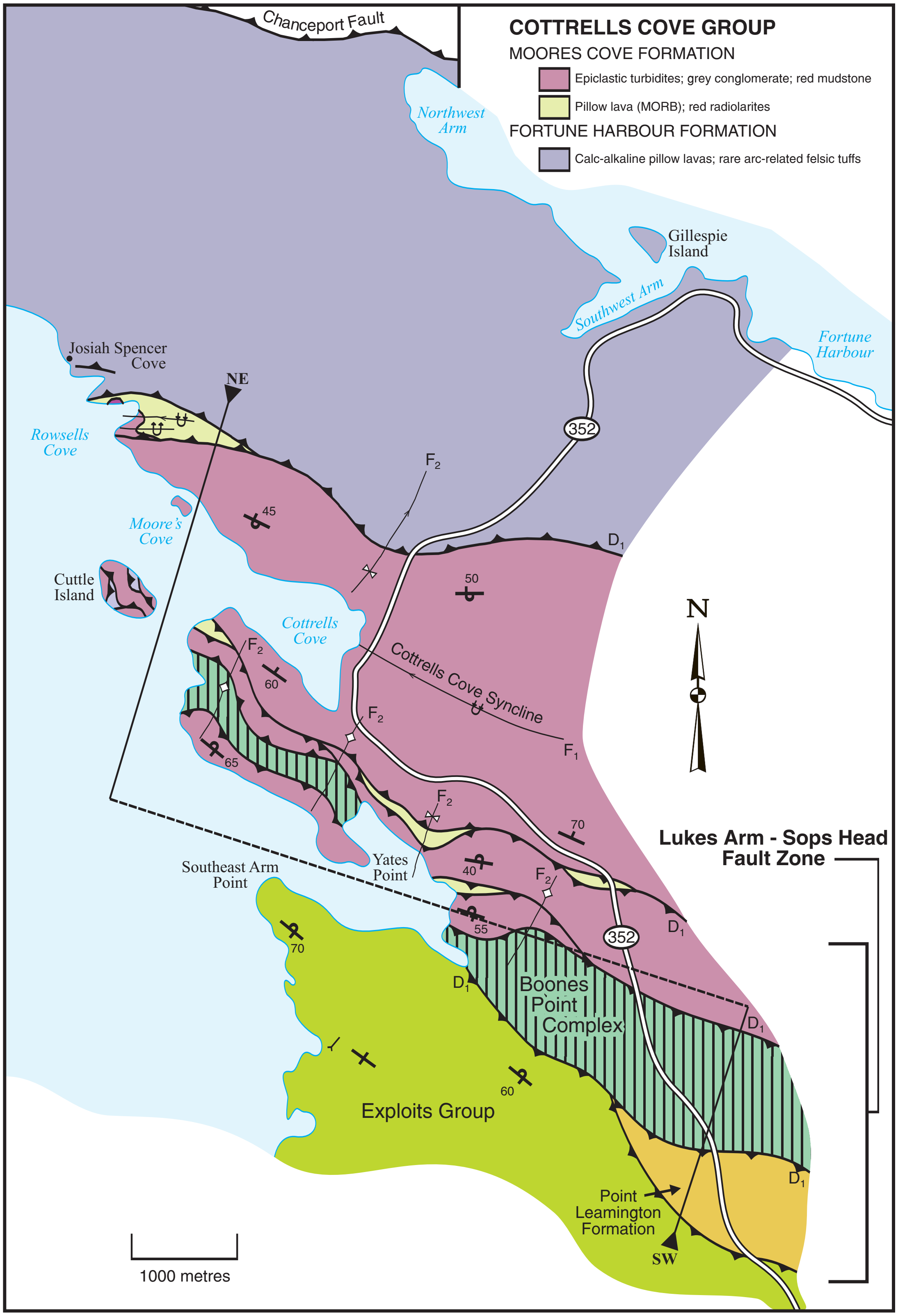


Figure 7 *Detailed lithostratigraphical and structural map of the type area of the Cottrells Cove Group adjacent to the Lukes Arm–Sops Head fault zone. Parts of the Exploits Group, the Point Leamington Formation and the Boones Point Complex are also illustrated. The Red Indian Line is locally coincident with the southwest margin of the Southeast Arm panel of the Boones Point Complex. Section line for Figure 31 is indicated.*

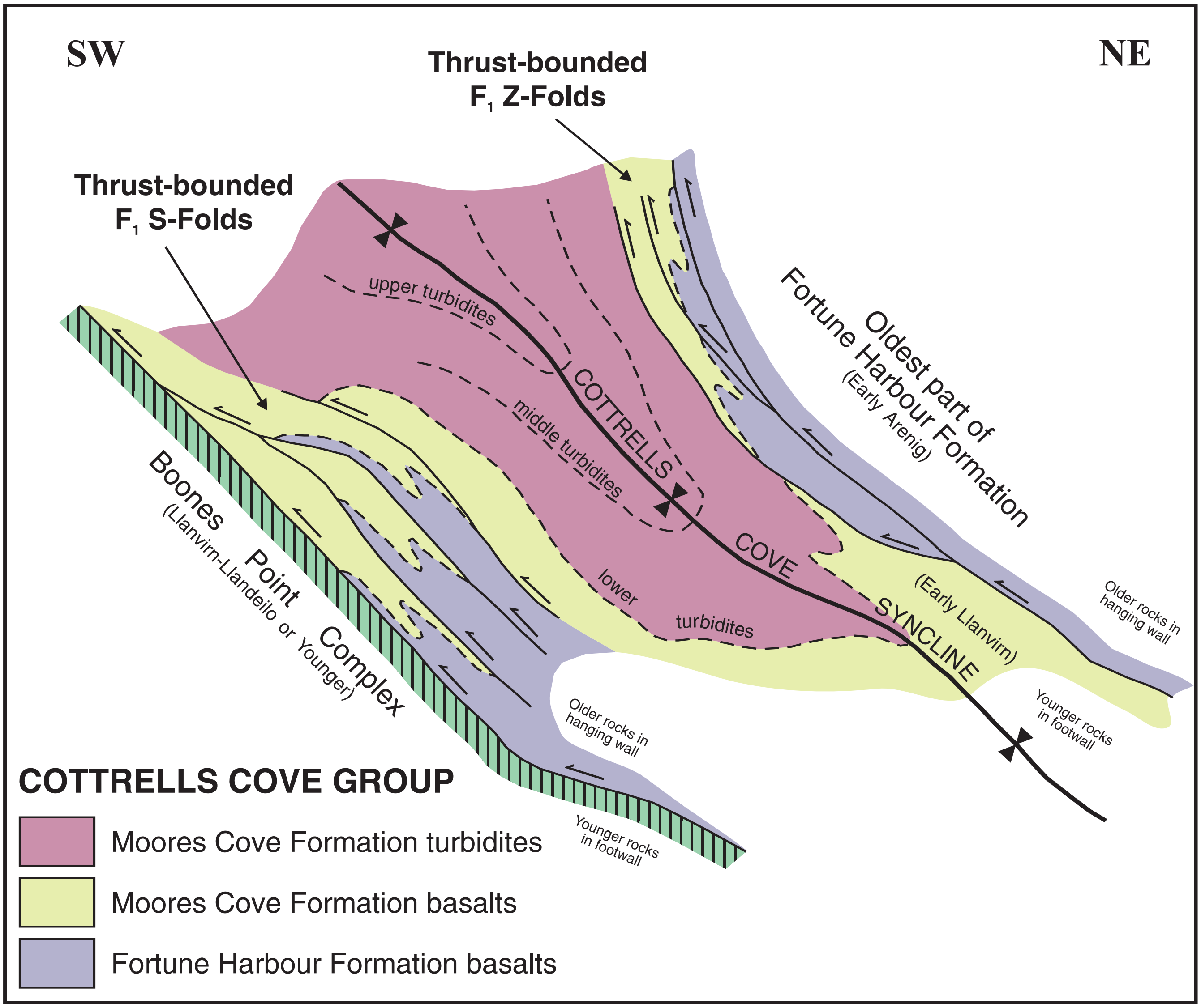


Figure 8. *Schematic cross section of the southwesterly overturned Cottrells Cove Syncline showing how the basal Moores Cove Formation is preferentially imbricated. On the opposing limbs of this regional fold, small thrust sheets contain parasitic F_1 folds that are bounded by D_1 faults. Asymmetrical Z-shaped F_1 folds occur on the inverted limb of the syncline; whereas, asymmetrical S-shaped F_1 folds are present on the right-way-up limb. Note that, on the southwest limb of the Cottrells Cove Syncline, the shear sense of thrust faults is different than the shear sense of the parasitic folds.*

rocks. Polymict conglomerate, distinguished by abundant limestone and rounded granite clasts, also forms highly conspicuous blocks in the melange tracts of the Boones Point Complex. Lithologically, they resemble conglomerate beds in certain late Early Ordovician horizons in the upper New Bay Formation, early Middle Ordovician horizons in the upper Moores Cove Formation, and late Early Silurian horizons in the Goldson Formation (Table 1).

In well-exposed coastal localities, primary debris flow deposits are preserved in large blocks within the block-in-matrix melange. The debris flow deposits contain a variety of reworked clastic sediments and thereby demonstrate a block-within-block hierarchy in the Boones Point Complex. Soft-sediment deformation textures in quick sands or chaotic muds are best preserved in the central parts of poorly-stratified sedimentary mega-blocks. These olistoliths commonly exceed 10 m in diameter and have thus been spared the strong regional deformation of the argillaceous matrix (O'Brien *et al*., 1994). However, near the Lukes Arm Fault, competent volcanic and granite blocks in the Boones Point Complex are structurally comminuted into yet smaller fragments and they are augened by a protoclastic foliation in the melange matrix (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65).

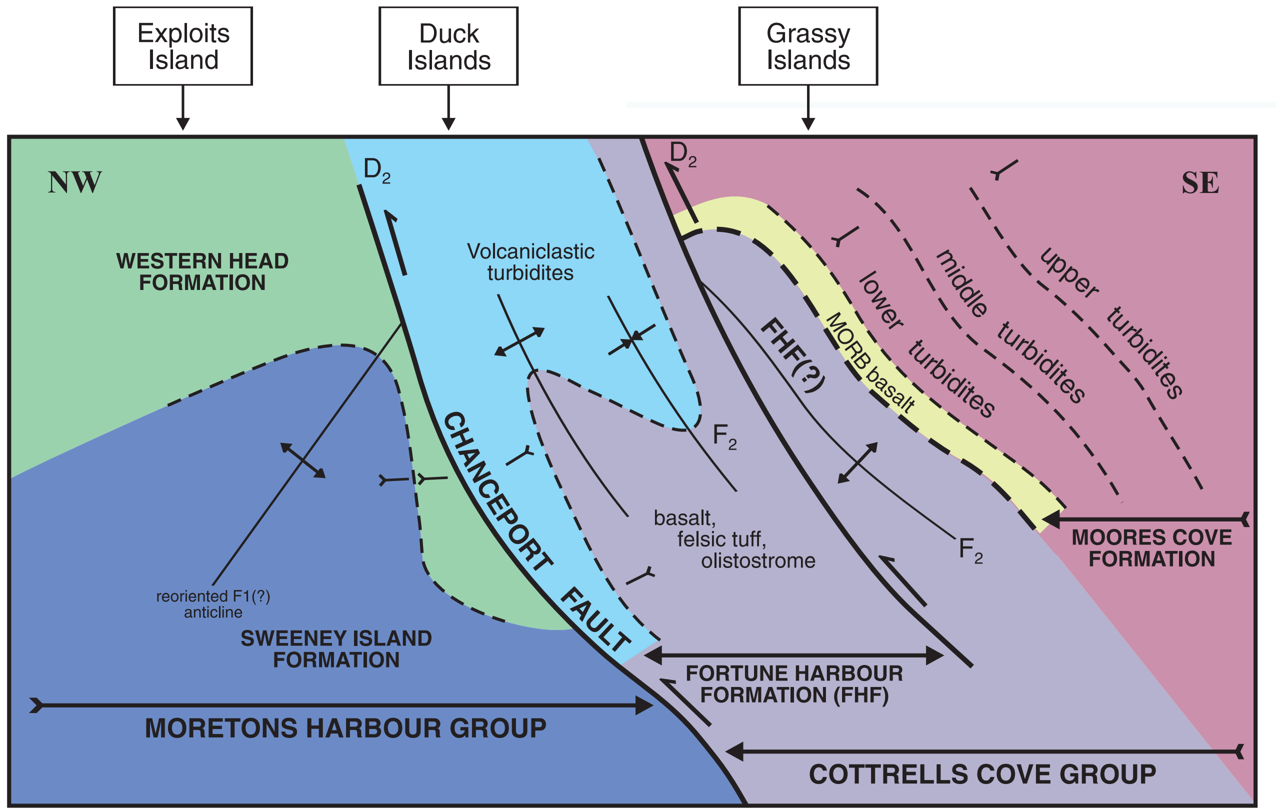


Figure 9. *Cross section, looking northeast, showing how southeast-dipping D_2 reverse faults displace F_2 folds within the Fortune Harbour and Moores Cove formations of the Cottrells Cove Group on the Duck Islands and Grassy Islands of the Bay of Exploits. Note that a Z-shaped F_2 fold affects the volcanic and sedimentary rocks of the Fortune Harbour Formation, which directly overthrusts the Western Head Formation of the Moretons Harbour Group on Exploits Island.*

The regionally deformed rocks of the Boones Point Complex occur in discrete fault-bounded tracts within an imbricate thrust zone developed near the Red Indian Line. Characteristically, individual tracts of block-in-matrix melange are discontinuous along strike, highly variable in size and lack internal stratification. The Boones Point Complex does not display continuous transitional zones from block-in-matrix melange to partially broken formations to mappable unbroken formations of volcanosedimentary strata. Furthermore, although the lithologies of most blocks in the melange can be generally matched with rock groups on the Fortune Harbour peninsula, they are commonly not found in the tectonically adjacent rock units in the imbricate thrust stack.

Structural Aspects

Along the southwest and southeast margins of the Boones Point Complex, tectonic melange forms crosscutting belts of block-in-matrix textured rocks. However, these belts typically measure a few decimetres to a few metres in maximum structural thickness (Plate 7). Tracts of tectonic melange were locally developed within the Boones Point Complex (and also near faults in adjacent Ordovician and Silurian map units) during several phases of thrusting, folding and cleavage development (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65).

Block-in-matrix melange of hard rock origin appears to have formed by tectonic shearing, structural comminution and physical disaggregation of strongly deformed strata (Plates 8 and 9). The protoliths of these high-strain rocks include the Lawrence Head Formation and Hummuck Island limestone of the Exploits Group, the Moores Cove Formation of the Cottrells Cove Group and the Sansom Formation of the Badger Group. Narrow belts of tectonic melange are mappable on 1:50 000 scale within the Lukes Arm fault zone on the Swan and Hornet islands (O'Brien, 1991b).

Block-in-matrix melange appears to have formed in strain-hardened tectonites during transient episodes of embrittlement, which were preceded and postdated by periods of ductile shearing in the vicinity of adjacent faults. In some ways, they are the upper crustal equivalents of the crosscutting veins of sheared pseudotachylite seen in deeper mylonite zones.

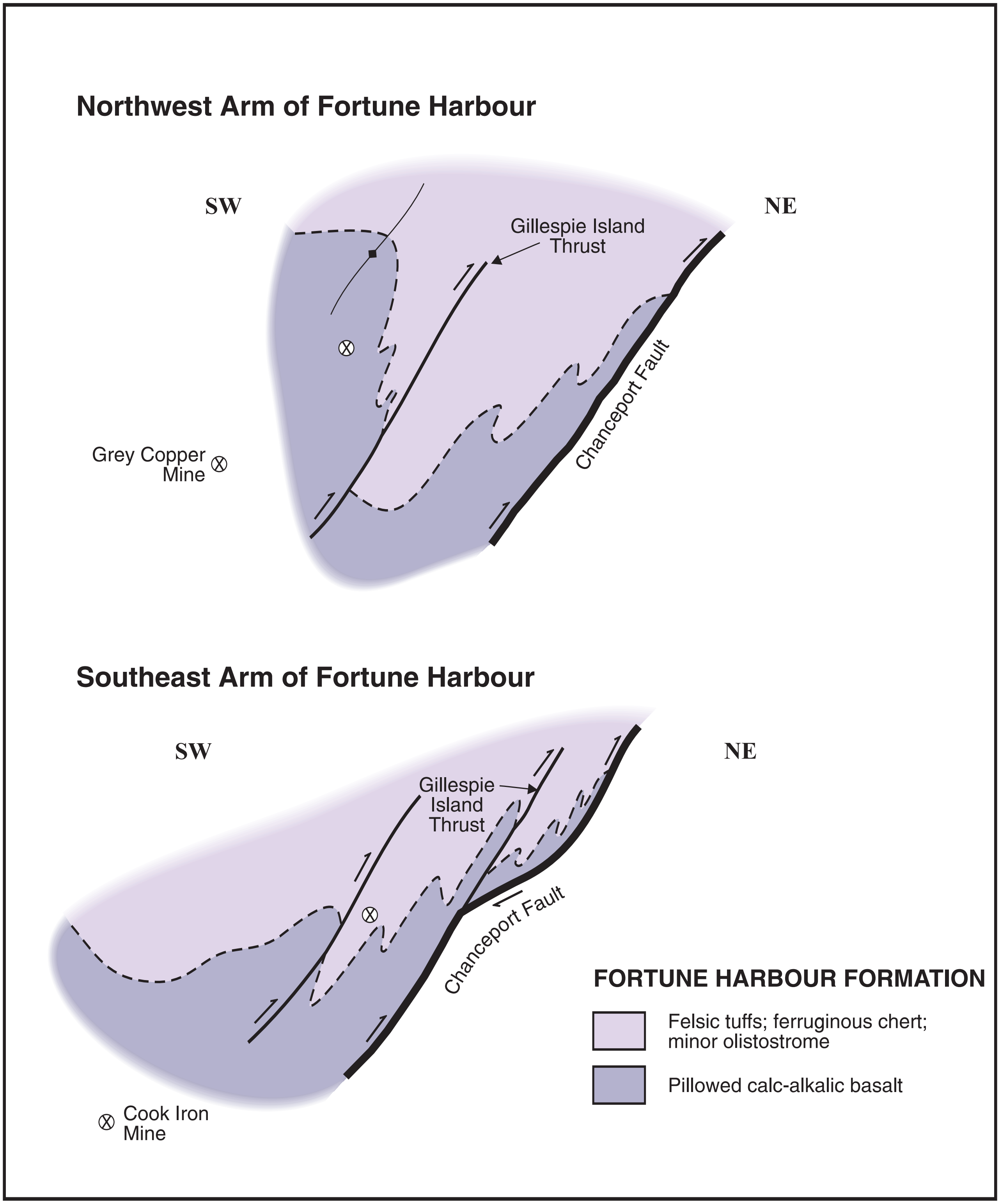


Figure 10. *Two cross sections of the type area of the Fortune Harbour Formation adjacent to the southwest-dipping Chanceport Fault. Northeasterly overturned trains of Z-shaped F_1 folds occur in separate thrust sheets on the flanks of synclines whose hinge zones are replaced by southwest-dipping D_1 thrust faults. Note how the tectonic panel of the Fortune Harbour Formation between the Gillespie Island thrust and the Chanceport thrust changes shape along strike as the bounding faults merge. The Grey Cooper Mine and the Cook Iron Mine lie, respectively, below and above the same stratigraphic horizon in the lower Fortune Harbour Formation but are situated in different tectonic panels.*

Plate 7. *A vertical cross-section through Moores Cove Formation turbidites shows sub-recumbent folds of a bedding-parallel foliation in the hangingwall of a small thrust fault within the Red Indian Line melange belt on northwestern Hornet Island. Note that individual beds were already detached and discontinuous prior to such folding, and that they were further disrupted and made podiform within the ductile fault zone located near the camera lens cap.*

Plate 8. *Well-foliated 'straightened' conglomerate of the New Bay Formation (formerly Rideout Point Formation) in north Campbellton (in background under hammer) passes into a structurally transgressive zone of matrix-foliated tectonic melange (in the foreground). The matrix contains locally derived blocks of randomly oriented 'straightened' conglomerate and porphyroclastic vein quartz.*

Age

Primary depositional laminations are locally preserved in the dark grey pyritiferous argillite which forms the sedimentary matrix of the block-in-matrix melange in the Boones Point Complex. However, extensive paleontological sampling of such laminae indicates that the melange matrix is unfossiliferous. This feature contrasts the mudstones and argillites in the Boones Point Complex with those in the adjacent Lawrence Harbour and Point Leamington formations (Table 1).

During the course of this study, searches were made for conodonts in several large limestone blocks within the melange; however, these investigations were unsuccessful. The Llanvirn–Llandeilo conodonts reported from melange blocks in the correlative Sops Head Complex (Nelson, 1981) remain the only known biostratigraphic control on the unit. They indicate a Middle Ordovician or younger age of melange formation.

Correlation

Petrochemical fingerprinting of outsized mafic volcanic blocks in the melange suggests that mafic lava flows in adjacent parts of the Moores Cove Formation of the Cottrells Cove Group and the Lawrence Head Formation of the Exploits Group were not the source of the exotic volcanic detritus in the Boones Point Complex (Dec and Swinden, 1994).

The Boones Point Complex is best correlated with olistostromal and tectonic melange tracts in the Sops Head

Complex. There, the pebbly mudstones and partially broken formations accumulated above relatively unbroken parts of the New Bay Formation of the Exploits Group (O'Brien, 2000). Based on regional considerations, it is possible that the Boones Point Complex represents a highly sheared tract of the Llanvirn–Llandeilo Dunnage Melange. In this regard, the Dunnage Melange is also seen to be fault imbricated with the Exploits and Badger groups on Upper Black Island, some 6 km southeast of Boones Point Complex exposures in the Little Grego Island area.

Plate 9. *A small shear zone in the Swan Island melange belt carries a strong matrix foliation and shows marked reduction of porphyroclast grain size. Many of the light-grey sedimentary blocks are discoidal and have asymmetric tails; whereas, some of the whiter tectonically rounded fragments may be vein quartz.*

Regional Interpretations

Nelson (1981) interpreted the enigmatic Boones Point Complex to have originated as the sedimentary infill of an Ashgill–Llandovery piggyback basin that was later fragmented and regionally deformed by a continuation of the same tectonic forces which initially formed the block-in-matrix olistostromal melange. An alternative view portraying the Boones Point Complex as a purely tectonic melange was put forward by LaFrance (1989). He reported that the block-in-matrix texture of the complex's melange units developed, for the first time, as a consequence of strike slip-dominated transpression during regional Late Silurian deformation. Blewett (1989, 1991) concluded that the Boones Point Complex was essentially a tectonized olistostrome, which was initially deformed in a regional folding and foliation-forming event during the Late Ordovician or the Early Silurian (Table 2). The complex was postulated to have been deformed again by more widespread Late Silurian structures, which affected Notre Dame Bay rock units as young as the Botwood and Springdale groups (Figures 3 and 5).

The main assumption regarding the age and origin of the Boones Point Complex has been that the ubiquitous, silicified, sandy wacke blocks which are broken down and distributed throughout the melange matrix belong, for the most part, to the Sansom Formation of the Exploits Subzone overstep sequence (Table 1). The view that Late Ordovician and Early Silurian olistostrome formation was triggered by thrust faults breaching the sea floor (Nelson, 1981) has been generally supported by the Cobbs Arm source of deformed marble blocks and the Llandovery depositional age of the Joey's Cove olistostromal melange in the Sansom Formation on New World Island (e.g., Reusch, 1987; Lafrance and Williams, 1992). Moreover, other Badger Group formations in central Notre Dame Bay that are not spatially associated with the Boones Point Complex display multiple debrite horizons carrying olistoliths of local sedimentary origin (e.g., Plate10), interstratified with coherent Ashgill turbidite successions. However, this interpretation does not consider that the sedimentary protolith for much of the melange in the Boones Point Complex was probably the Middle Ordovician Moores Cove Formation (e.g., Plate 11) and that the region's known Late Ordovician olistostromes are notably orders of magnitude smaller and thinner than those of the Boones Point Complex.

EXPLOITS SUBZONE

Exploits Group

The Exploits Group is the most extensive lithostratigraphic unit of the Exploits Subzone in the central Notre Dame Bay region (Figure 3). The well-exposed type area is located in the New Bay–Thwart Island–Lewisporte area (Kean *et al*., 1981).

Distribution and Thickness

The Early Paleozoic Exploits Group crops out over a 600 km^2 area and is estimated to be approximately 3.5 km thick (O'Brien *et al*., 1997). Two main structural tracts are identifiable within the Exploits Group. To the southwest, in a southeast-plunging fault-modified synclinorium, the regional stratigraphic facing direction of formations is southwestward (Figure 11). To the northeast, in a doubly plunging M-shaped anticlinorium, the regional stratigraphic facing direction of formations is northeastward. Unit thickness in the middle part of the Exploits Group appears to increase northeastward.

Plate 10. *A mud-matrix olistostrome belonging to the* D. anceps *Zone of the Ashgillian Point Leamington Formation contains older blocks derived from the lower bioturbated chert unit (below camera lens) and younger Caradoc black siltstone unit (right of camera lens) of the stratigraphically lower Lawrence Harbour Formation.*

In the north, the Lukes Arm Fault separates the upper part of the Exploits Group from the upper part of the Cottrells Cove Group (Figure 11). In the south, the New Bay Fault separates the middle part of the Exploits Group from the upper part of the Wild Bight Group.

Stratigraphic Nomenclature

The formally defined lithostratigraphic units in current use were established by Helwig (1969), although a part of the region underlain by the Exploits Group was originally mapped and subdivided by Heyl (1936). As first defined, the term Exploits Group was employed to include late Middle and Late Ordovician formations of Caradoc shale and Ashgill greywacke.

Helwig (1967, 1969) originally separated the pre-Caradoc portion of the Exploits Group into four conformable units of formational rank. From oldest to youngest, these regionally mappable units are the Tea Arm Formation, the Saunders Cove Formation, the New Bay Formation and the Lawrence Head Formation. During the mapping for this report, two informal units–the Strong Island chert and the Hummock Island limestone–were recognized in the uppermost (pre-Caradoc) part of the Exploits Group.

In some areas, correlatives of the upper New Bay Formation, the Lawrence Head Formation and the Strong Island chert have been given a variety of other local names, or have been assigned to units other than the Exploits Group [e.g., Kay's (1975) Campbellton Sequence; Table 1].

Stratigraphic Revision

Dean (1977) informally dropped the Saunders Cove Formation from the Exploits Group but reaffirmed the stratigraphic order of the remaining three formations. The name Saunders Cove Formation is retained in this study and its outcrop area has been expanded. Present investigations indicate that the New Bay Formation, the Lawrence Head Formation and the Strong Island chert are widely distributed but that they are laterally discontinuous in certain localities (O'Brien *et al*., 1997). Regionally mappable units of Caradoc shale and Ashgill greywacke have been removed from the Exploits Group, following the recommendation of Kean *et al*. (1981).

Lithostratigraphy

For the purposes of this report, the internal map units deemed to comprise the Exploits Group are, in ascending order, the Tea Arm Formation, the Saunders Cove Formation, the New Bay Formation, the Lawrence Head Formation, the Strong Island chert and the Hummock Island limestone (Figure 12). In the study area, several formations have been further separated into members (Figure 12) that can be portrayed on 1:50 000 scale geological maps of the Exploits Group (O'Brien, 1990, 1992, 1993).

Though the stratigraphic base of the Exploits Group is nowhere exposed, the stratigraphic top of the group is observable in several localities beneath the Llandeilo–Caradoc Lawrence Harbour Formation. The Strong Island chert commonly underlies the Lawrence Harbour Formation, where the uppermost part of the Exploits Group is preserved and stratigraphically intact. In many but not all primary contact locations, the spectacular pillow lavas of the Lawrence Head Formation overlie the turbidites of the New Bay Formation and gradationally underlie the Strong Island chert. In areas where the Lawrence Head Formation was not deposited, the Strong Island chert directly overlies the New Bay Formation.

Helwig (1969) thought that the early–middle Paleozoic formations of the Exploits Group represented a paraconformable succession of deep-marine strata. The writer would concur that all Ordovician map units in the Exploits

Group and the Exploits Subzone overstep sequence are observed to be in conformable contact in one or another part of the area surveyed.

Plate 11. *Asymmetric shear bands in a tectonic melange on Swan Island deform an anastamosing protoclastic foliation that formed in highly deformed rocks of the Moores Cove Formation. Note that two structural domains, one to the left and the other to the right of the shear bands, illustrate a considerable variation in the spacing of the foliation seams (dark-coloured areas) and the relict size of the metasedimentary lithons (light-coloured areas).*

Lithology and Stratigraphy of the Tea Arm Formation

The Tea Arm Formation of the Exploits Group is well exposed near Strong Island Sound on the peninsula separating the Southwest Arm and South Arm of New Bay (Figure 13). There, the minimum stratigraphic thickness of the formation is estimated at about 1 km. The volcanic rocks of the Tea Arm Formation have been lithostratigraphically subdivided into three divisions (O'Brien, 1990), which were revised and informally named by O'Brien *et al*. (1997). In ascending order, they are the Little Arm East member, the Pushthrough member and the Pleasantview member (Figure 12).

Little Arm East Member

The stratigraphically lowest Little Arm East member crops out in the faulted core of the Tea Arm Anticline. It is composed of highly vesicular, variably porphyritic, poorly-stratified basalt flows and rare vitric mafic extrusions. These are intruded by large gabbro sills and swarms of diabase dykes. Lava tubes and pillows are separated by interstitial chert; however, major horizons of interstratified sedimentary rocks are absent.

Pushthrough Member

The gradationally overlying Pushthrough member of the Tea Arm Formation is relatively thin (ca. 75 m or less) and is largely made up of rhyolitic tuff and mixed felsic–mafic breccias. Banded quartz-feldspar crystal tuffs are locally intruded by diabases that were disrupted during pyroclastic flow. Coarse agglomerates carry felsic and mafic bombs and are laterally transitional with the finer grained pyroclastic rocks. In Tea Arm bottom, siliceous iron formation is present as a thin red cap rock to these submarine volcanic deposits, although it is more commonly found as blocks within felsic breccias.

Pleasantview Member

The youngest division of the Tea Arm Formation, the Pleasantview member, directly overlies Pushthrough felsic tuffs. Near its stratigraphic base, several basalt flows contain pillows that are collapsed, drained, sheared and altered. All of these events occurred prior to the injection of diabase dykelets, which occupy conduits that are also filled by variegated cherts. At this stratigraphic position in the Tea Arm Formation, limestone deposition was most extensive, punctuating some of the intervals between the collapse and shearing of pillows and renewed basalt eruption. Higher in the succession, rhythmically stratified intervals predominate in the Pleasantview member. Most of the upper part of this division is represented by green pillow lava grading to red hyaloclastite, and then to volcaniclastic wacke and grey or red, ripple-marked cherty argillite.

Geochemistry of the Tea Arm Formation

One of the significant characteristics of the Tea Arm Formation is the range of distinctive geochemical signatures for its basalts and rhyolites (Dec *et al*., 1992; O'Brien *et al*., 1997). Subalkaline pillow lavas displaying light rare earth element (LREE) patterns with prominent negative Nb and Ta and positive Th anomalies occur in all members of the formation. The total abundance of the LREE elements is generally low and similar to the island arc tholeiites and calc-alkaline basalts found in primitive intraoceanic arcs.

Felsic volcanic rocks in the Pushthrough member, dated as late Tremadoc–early Arenig in age, are high-silica trondhjemitic rhyolites. They are associated with strongly LREE-depleted basalt flows in the lower Pleasantview member. This indicates that the magma batches feeding such lavas were not melted from a source with a normal mantle composition but from one that had been metasomatized by fluids from a fossil subduction zone. The trondhjemitic rhyolites and island arc tholeiites near the Tea Arm massive sulphide prospect are very similar to those found in the Wild

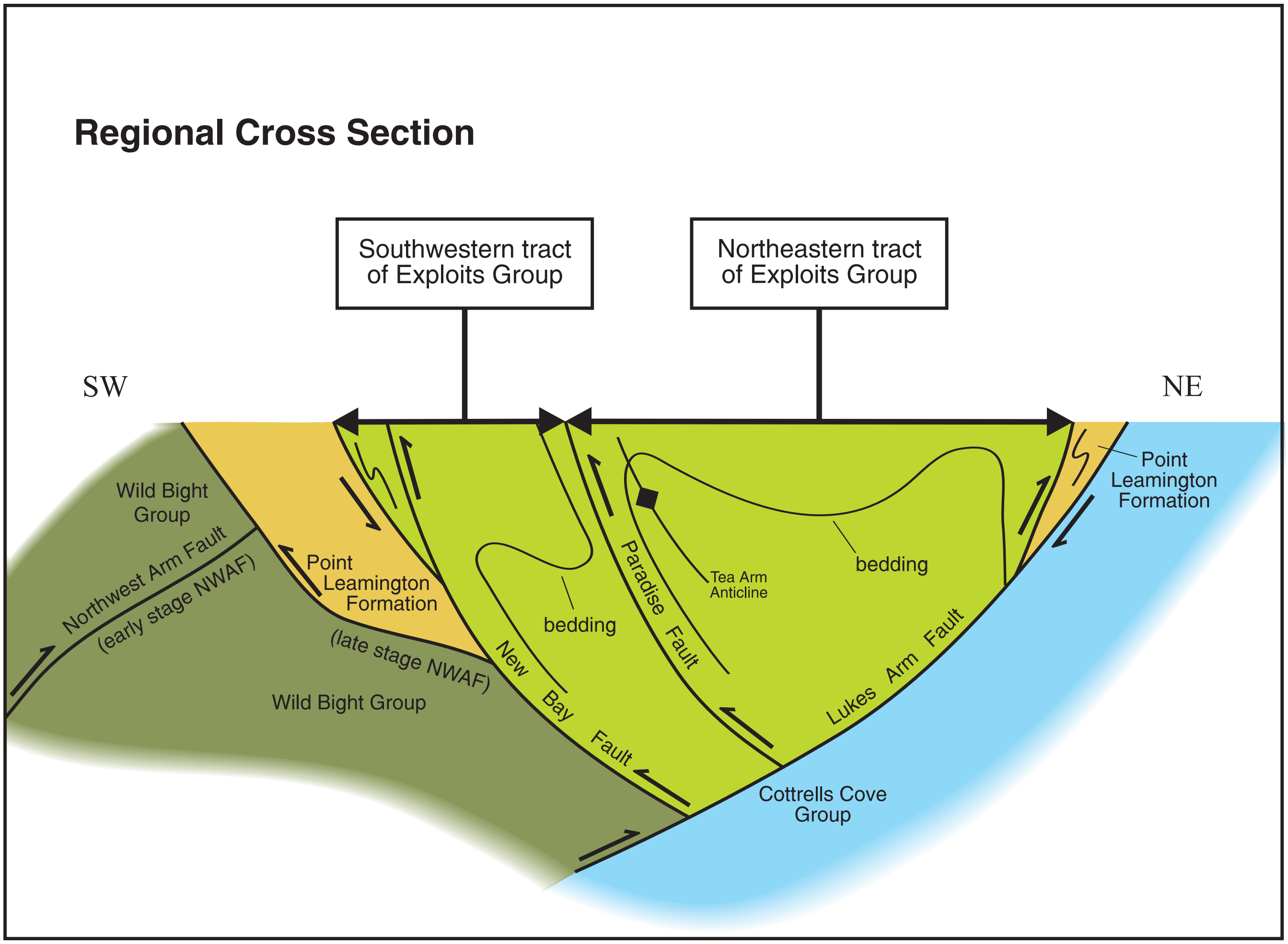


Figure 11. *Regional northeast–southwest cross section showing some of the major D_1 thrusts and F_1 folds affecting the northeastern and southwestern structural tracts of the Exploits Group. Also shown are some of the southwest-dipping and northeast-dipping D_1 faults that juxtapose the Exploits Group with the Cottrells Cove Group, the Wild Bight Group and the Point Leamington Formation. The faults illustrated were originally mapped and named by Helwig (1967).*

Bight Group near the Lockport massive sulphide prospect (Swinden, 1987; MacLachlan, 1998).

Lithology and Stratigraphy of the Saunders Cove Formation

The Saunders Cove Formation of the Exploits Group crops out near the South Arm of New Bay in proximity to the Paradise Fault and occurs in both the northeastern and southwestern tracts of the Exploits Group (Figures 3 and 13). The maximum stratigraphic thickness of this formation is about 400 m.

The Saunders Cove Formation is largely composed of red chert interbedded with red and green siliceous argillite. Some cherts display traction-produced sedimentary structures and are graded from siltstone to mudstone. They presumably represent original clastic beds that are now replaced and altered to chert. Other massive, amorphous, conchoidal and finely laminated varieties contain recrystallized microspheres that are possibly relics of radiolaria. These could be biogenic cherts or chemical precipitates (Swinden, 1976).

Extensive iron oxidation, collapse and chertification of strongly LREE-depleted pillow lavas occurs at the base of the Saunders Cove Formation (Plate 12). Up-section, the red chert and iron formation is host to a minor but generally increasing amount of coarser clastic sediment. Massive, 5 to 15 m thick turbidite units are graded from a basal lag composed of polymict conglomerate to an upper interval made up of pebbly to sandy feldspathic wacke. A distinctive laminated to thinly-bedded marker horizon of dark grey shale and light grey carbonaceous argillite is approximately 25 m thick and is traceable for approximately 6 km along strike in the Paradise–Little Arm East area.

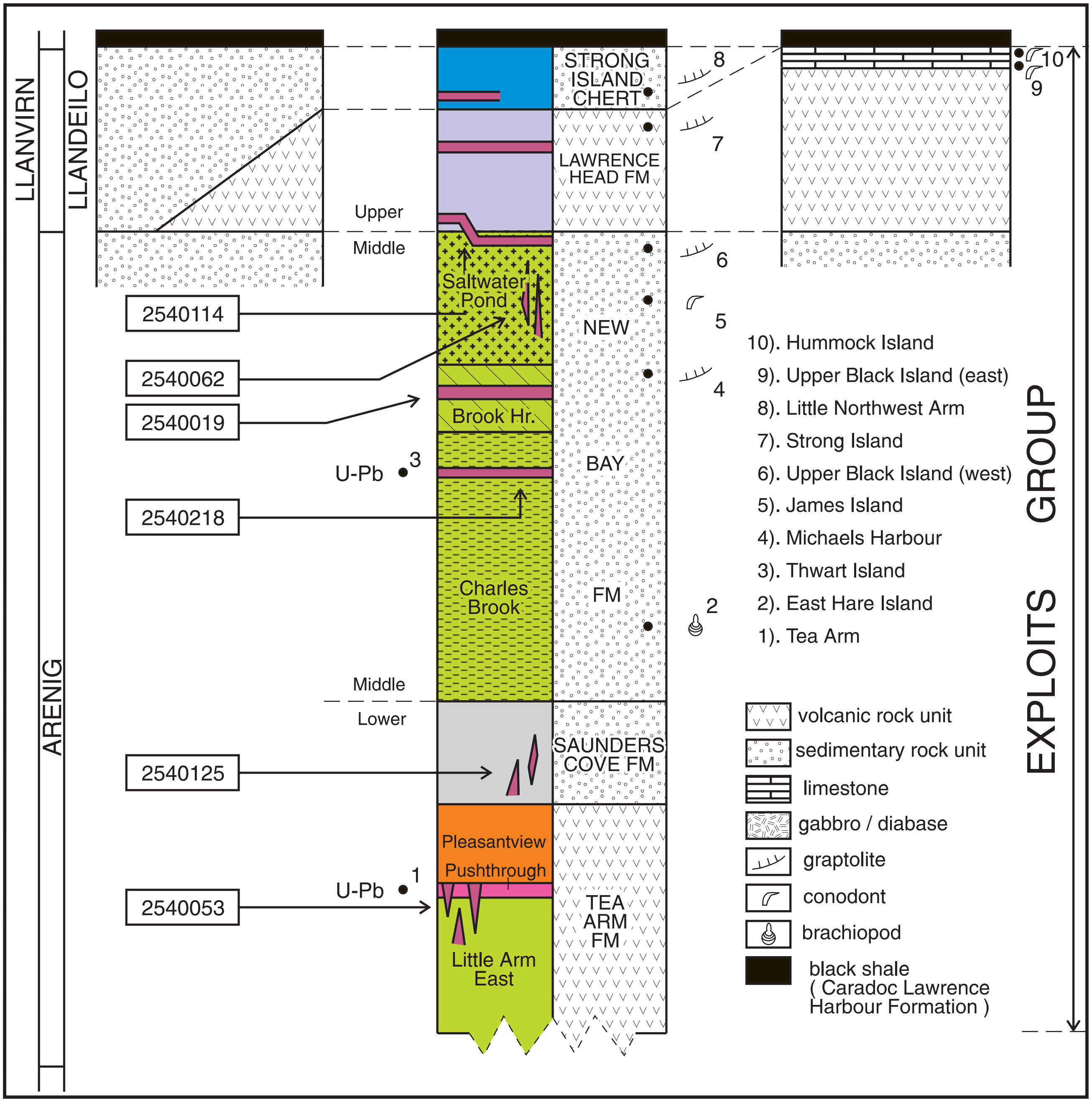


Figure 12. *Generalized stratigraphic column of formal and informal rock units in the Exploits Group of known Arenig, Llanvirn and Llandeilo age. Positions of fossil localities, U/Pb-dated rocks and minor intrusions with geochemical data are indicated (*see *O'Brien* et al., *1997).*

Lithology and Stratigraphy of the New Bay Formation

The New Bay Formation of the Exploits Group generally consists of deep-sea turbidites (Helwig, 1967) which locally attain a maximum stratigraphic thickness in the order of 1500 to 2000 m. In ascending order, three constituent members are named the Charles Brook member, the Brook Harbour member and the Saltwater Pond member (Figure 12).

Charles Brook Member

The oldest Charles Brook member of the New Bay Formation is generally composed of sandy and silty turbidites and is approximately 850 m thick. These strata show complete or nearly complete Bouma intervals and display abundant flutes, tool marks and other bottom structures. In the lower part of the Charles Brook member, the interbedded light-grey siltstone and grey sandstone succession is scoured by massive pebbly wackes with rare conglomerate lags.

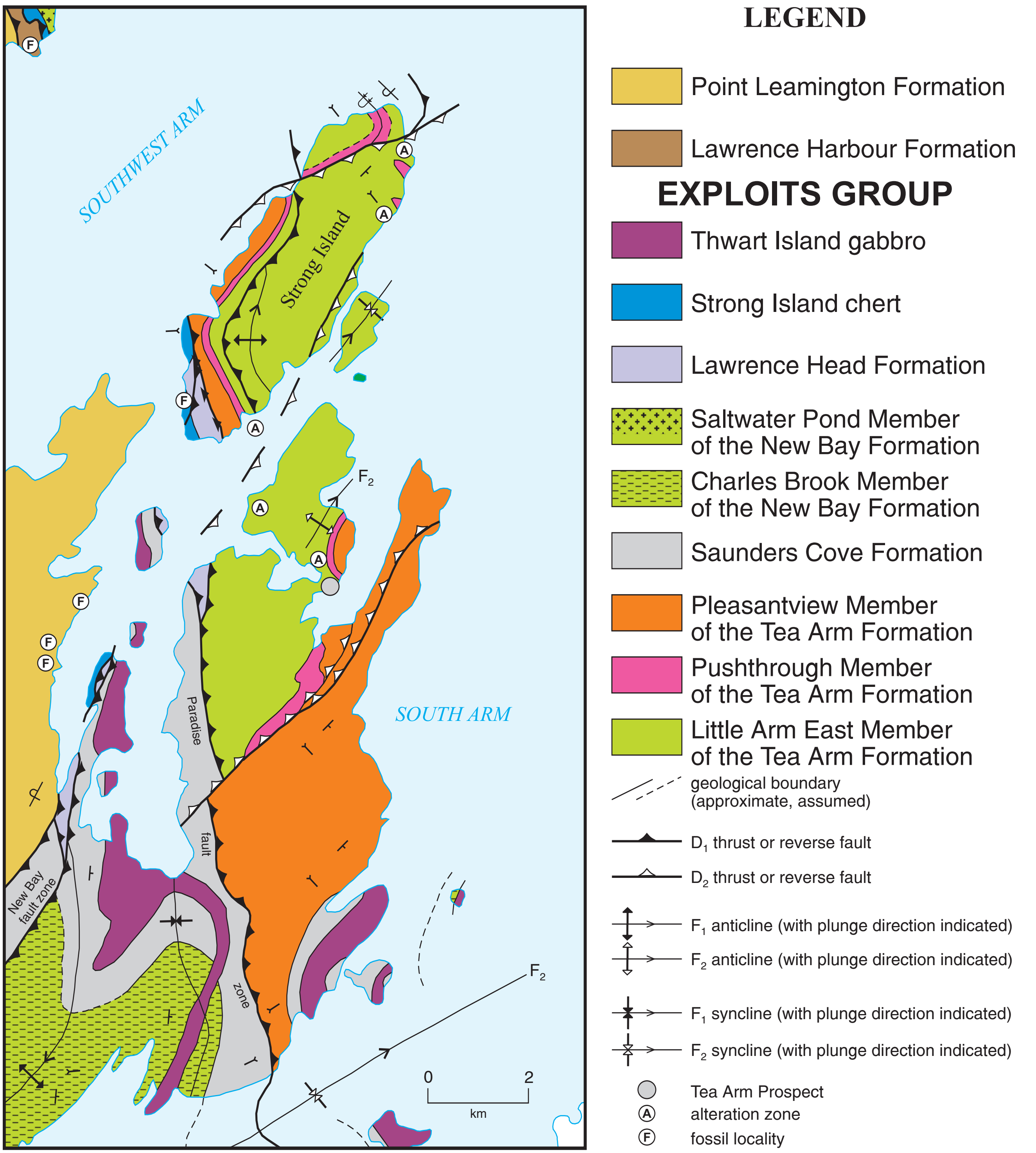


Figure 13. *Geological map of the Exploits Group in the vicinity of the Tea Arm prospect emphasizing the structural disposition of the stratigraphic members of the Tea Arm Formation. Positions of other alteration zones with pyrite or chalcopyrite stringers are also indicated.*

Such thick, coarse-grained, siliciclastic deposits are rich in basaltic and dioritic epiclasts as well as sandstone intraclasts. Conspicuous lapilli stone horizons, rich in andesitic tephra (O'Brien *et al*., 1994), are also present at several horizons in the Charles Brook member.

Numerous slump-folded intervals of intrabasinal siltstone and sandstone occur in the upper part of the Charles Brook member. In certain stratigraphic horizons, the upper parts of slump sheets are observed to be partially disaggregated on the limbs of soft-sediment fold nappes. In places,

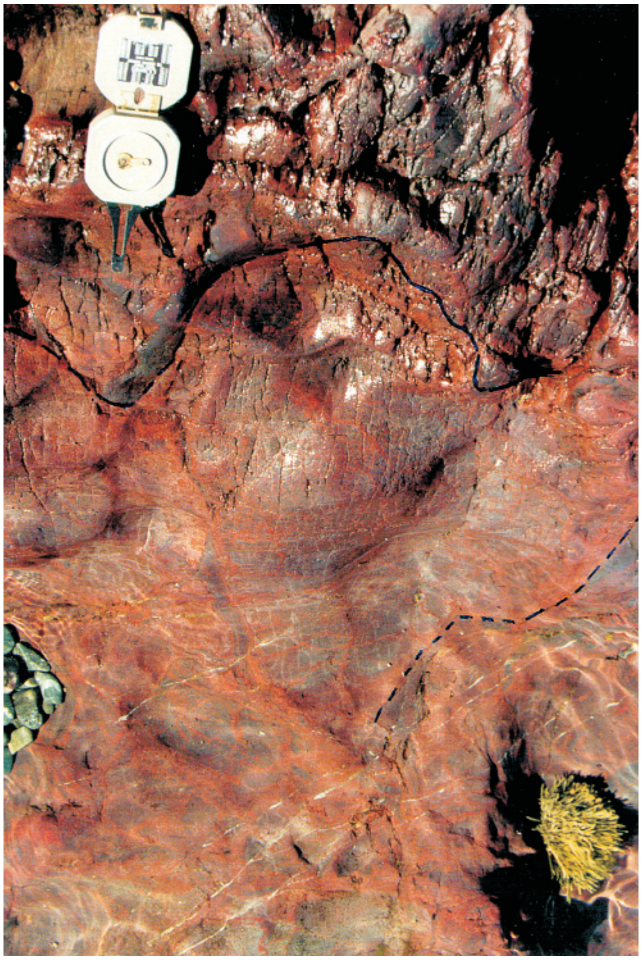


Plate 12. *Hematized and silicified basalt preserved within an oxide-facies iron formation at the base of the Saunders Cove Formation of the Exploits Group near Saunders Cove. Note the relict shape of partially collapsed and netveined pillow lava.*

the inverted and right-way-up limbs of recumbent slump folds were eroded by overlying sandstones. In other localities, sills and dykes of sandstone were injected upwards in a complex fashion from discrete quicksand beds (Plate 13). These events may have occurred during subsidence and compaction of a partially lithified sequence of Charles Brook strata.

Brook Harbour Member

The intermediate Brook Harbour member of the New Bay Formation is dominated by dark grey siltstone and dark-grey to black mudstone and is approximately 350 m thick. Subordinate turbidites are thin-bedded, light grey, faintly graded sandstones that illustrate only partial Bouma intervals. These sandstones comprise considerably less of the total thickness of the Brook Harbour member than do sandstone turbidites lower in the New Bay Formation. In several different localities around the Bay of Exploits, fragmentary pieces of poorly preserved graptolites have been identified in the Brook Harbour member.

Saltwater Pond Member

The youngest Saltwater Pond member of the New Bay Formation contains silicified grey argillites and medium-bedded siltstone turbidites low in the section and poorly sorted granular epiclastic wacke, red and green siliceous argillite and minor chert high in the section. In total, this member is approximately 500 m thick.

The intervening sequence of the Saltwater Pond member is marked by thick red- and grey-coloured olistostromes, grey debrites containing partially detached slump sheets and discrete sedimentary blocks, and graded polymict conglomerate having a variety of exotic granitoid and limestone clasts. The most conspicuous olistostromes possess blocks of resedimented well-rounded boulder conglomerate (Plate 14).

Lithology and Stratigraphy of the Lawrence Head Formation

The Lawrence Head Formation is a widespread sporadically developed unit that is discontinuously exposed around the periphery of the outcrop area of the Exploits Group. It occurs between Point Leamington and Little Northwest Arm in the west, between Cottrells Cove and Lawrence Harbour in the north and between Purbeck Cove and Campbellton in the south (O'Brien *et al*., 1997). The maximum stratigraphic thickness of the Lawrence Head Formation is about 400 m.

Pillow lavas of the Lawrence Head Formation (Plate 15) are intruded by columnar-jointed diorite sills linked together by diorite dykes. Many of these hypabyssal rocks pass gradationally into marginal zones of highly vesicular diabase, whose offshoots are observed to feed lava tubes. Basalts in the upper part of the Lawrence Head Formation are interstratified with and gradationally overlain by green (or red) ferruginous chert and subordinate volcaniclastic wacke. However, in localities where these cherts are thin or were never deposited, limestone lenses are intercalated with mafic lavas in the uppermost part of the formation. Felsic pyroclastic rocks have not been observed within the Lawrence Head Formation, although very thin shard-rich tuff beds are locally conspicuous in the conformably underlying turbidites of the New Bay Formation.

In the Lewisporte region, rocks similar to those in the Lawrence Head Formation occur in the upper part of the Loon Harbour volcanic formation (Plate 16; Table 1). There, pillow lava and hyaloclastite underlie a variably thick unit of variegated chert, wacke and minor carbonate, which is gradational upward into graptolite-bearing Caradocian strata (O'Brien, 1992). On the basis of its within-plate geochemical signature and abundant diorite sills, the westernmost outcrop of the Loon Harbour volcanic formation is considered

Plate 13. *Coarse-grained quicksands in argillaceous strata of the Charles Brook member of the New Bay Formation near East Hare Island contain fragments of the original muddy wallrock of the intrusive bodies. Note how some of the wet-sediment intrusions are highly curviplanar (bending continously from dyke to sill orientations). Also note how some quicksand sills crosscut by sandstone bodies generate new sandstone dykes, possibly indicating a complex relationship between compaction and hydraulic fracturing. End of hammer points toward the stratigraphical top of the sedimentary section.*

Plate 14. *Olistostrome within the Saltwater Pond member of the New Bay Formation near Winter Tickle displays a red mudstone matrix with outsized blocks of polymict boulder conglomerate. Smaller rounded to angular clasts of granite and vesicular basalt occur in a slump-folded, chaotically deformed matrix.*

as an easterly inlier of the Lawrence Head Formation of the Exploits Group (O'Brien *et al*., 1997).

Lithology and Stratigraphy of the Strong Island Chert

The Strong Island chert is a regionally widespread unit without formal rank that occurs throughout the Exploits Group and in the eastern Wild Bight Group (O'Brien, 1990). The stratigraphic thickness of the unit varies from 0 to 400 m thick in the Exploits Group.

The Strong Island chert is observed to stratigraphically overlie the sedimentary rocks of the New Bay Formation and the volcanic rocks of the Lawrence Head Formation (O'Brien, 1991a; Dec *et al*., 1992). However, in Little Arm East and on Strong Island , it is directly juxtaposed against the Tea Arm and Saunders Cove formations of the Exploits Group (Figure 13). The Strong Island chert conformably underlies a grey bioturbated chert near the base of the Lawrence Harbour Formation (Table 1).

The Strong Island chert is mostly composed of grey ribbon chert interbedded with graded volcaniclastic wacke (Plate 17). Certain ribbon cherts are thought to contain poorly preserved radiolaria (Dec *et al*., 1992). Some green lithic wackes in the Strong Island chert display Bouma intervals and are particularly rich in aphanitic mafic volcanic clasts. Others display conspicuous, oblong, round-edged fragments of intraformational black siliceous argillite. In some areas, decimetre-thick wackes and small debris flows have fresh detrital plagioclase feldspars, but also crystalline quartz and some coarse-grained phyllosilicates (Dec *et al*., 1992).

Plate 15. *Spectacular pillowed lavas of the Lawrence Head Formation in its type area, viewer looking at about 1000 m² of the top surface of a steeply dipping stratification plane. Overlying chert beds occur to the right of the scree for some 20 m above sea level.*

Interbedded, red and green, banded cherts dominate the succession of the Strong Island chert where the map unit is thin. Such chert beds are notably ferruginous where they are in direct stratigraphic contact with pillowed basalts and pillow breccias. Turquoise cherts are interstratified with subalkalic to alkalic pillow lavas in the transition zone between the Lawrence Head Formation and the Strong Island chert (O'Brien *et al*., 1997).

The middle to upper part of the Strong Island chert is commonly composed of a succession of altered grey sedimentary rocks (Plate 18). These include thin-bedded, mottled cryptocrystalline argillites and medium-bedded, coarse-grained, carbonate-altered siliciclastic wackes. In places, fine-grained sandstone turbidites within this grey bed succession display internal Bouma intervals that were distorted, partially re-amalgamated, silicified and then variegated. Such strata typically occur well up-section from any interstratified pillow lavas. The grey replacement cherts, which appear to be best developed where the Strong Island chert is thickest (about 350 to 400 m thick), dominate in areas where the Lawrence Head Formation is locally missing from the Exploits Group.

The Strong Island chert is the youngest unit of the Exploits Group to host the regionally developed suite of Thwart Island gabbro laccoliths (Plates 19 and 20; Figure 12; O'Brien *et al*., 1997).

Lithology and Stratigraphy of the Hummock Island Limestone

The Hummock Island limestone is restricted to the northeastern part of the Exploits Group and is exposed on islands in the western Bay of Exploits (O'Brien *et al*., 1997). It varies from 1 to 5 m in total thickness and is developed in locations where the ferruginous beds of the Strong Island Chert are very thin or completely absent. The Hummock Island limestone has been assigned to the Exploits Group, although its limited regional extent precludes a formal stratigraphic designation.

In places where primary stratigraphic boundaries are preserved, the Hummock Island limestone occurs above within-plate (LREE-enriched) tholeiites of the Lawrence Head Formation (Plate 21) and below the grey bioturbated chert near the Llandeilo–Caradoc Lawrence Harbour Formation. However, the Hummock Island carbonates are most commonly situated in thrust sheets detached from surrounding formations. Within mylonite zones near the Lukes Arm Fault, the unit is highly sheared and metamorphosed to marble, and has potential as a tidewater- accessible industrial mineral prospect (Dean, 1978).

Conodonts and shelly faunas collected from the Hummock Island limestone indicate an age range that predates and partially overlaps the age range of the Cobbs Arm limestone of New World Island (Figure 3; Table 1; section on Age, page 34).

Stratigraphic Summary of the Exploits Group

The Tea Arm Formation of the Exploits Group consists of a lower subunit of pillow lava with a mafic dyke swarm, an intermediate subunit of felsic pyroclastic rocks, and an upper subunit of pillow lava, pillow breccia, interflow limestone and rare laminated argillite (O'Brien, 1990; Figure 12). The Saunders Cove Formation sharply overlies the primitive arc basalts of the Tea Arm Formation and, in its lower part, includes an oxide facies iron formation associated with red chert, red argillite and a distinctive dark-grey laminated shale. Together with the Saunders Cove Formation, the Little Arm East, Pushthrough and Pleasantview members of the Tea Arm Formation comprise the lower Exploits Group.

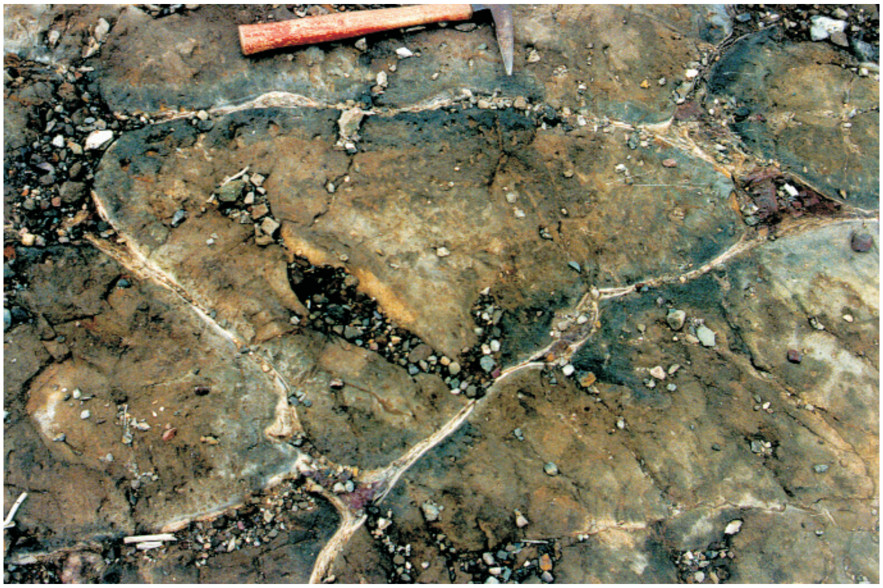


Plate 16. *Cuspate pillow lavas with interstitial red chert, outcropping to the east of Lewisporte Harbour, belong to the Lawrence Head Formation of the Exploits Group (Loon Harbour volcanics of the Campbellton Sequence). Stratigraphical top of the basalt lavas is toward the geological hammer.*

Conformably overlying the Saunders Cove Formation is the oldest of three, stratigraphically gradational subdivisions of the turbidite-dominated New Bay Formation (O'Brien, 1993). The oldest subunit is generally sandy and medium bedded, the intermediate subunit is mostly muddy and thin bedded, and the upper subunit is typically conglomeratic and thick bedded. The Charles Brook, Brook Harbour and Saltwater Pond members of the New Bay Formation comprise the middle Exploits Group.

The overlying Lawrence Head Formation contains columnar-jointed diorite sills that intrude pillow lavas and pillow breccias, which are lithologically similar but geochemically distinct from those in the older Tea Arm Formation. Within-plate basalts at the base of the Lawrence Head Formation are interstratified with graptolite-bearing turbidites, while those at top of the formation are interbedded with graptolite-bearing cherts. The succeeding Strong Island chert represents a relatively condensed, ferruginous hemipelagic succession dominated by biogenic and replacement chert but also composed of subordinate amounts of volcaniclastic and polymictic debrite and pillowed basalt (Dec *et al*., 1992).The Hummock Island limestone replaces the Strong Island chert in the northeastern part of the group. Together, the Lawrence Head Formation, the Strong Island chert and the Hummock Island limestone comprise the upper Exploits Group.

Age

A quartz-feldspar crystal tuff from the type area of the Pushthrough member in Tea Arm has ca. 486 Ma zircons, indicating a latest Tremadoc–early Arenig age of eruption (O'Brien *et al*., 1997). Crinoid stems have been previously recognized in limestones of the Pleasantview member of the Tea Arm Formation; however, biostratigraphic age assignments were not reported (Helwig, 1967).

In two separate fossil localities near East Hare Island, the Charles Brook member of the New Bay Formation has yielded undated inarticulate brachiopods (Figure 12). The mudstones of the Brook Harbour member were found to contain abundant fragmentary (but unidentifiable) graptolites at fossil localities near Phillips Head, Thwart Island and Michaels Harbour. However, on James Island, detrital limestone clasts in the Saltwater Pond member have been reported to yield early Arenig conodonts (Hibbard *et al*., 1977). These provide the only known older age limit for deposition of the New Bay Formation.

Graptolites from the *Undulograptus austrodentatus* Zone were recovered from the upper Saltwater Pond member of the New Bay Formation, the upper Lawrence Head Formation and the lower Strong Island chert (Williams *et al*., 1992; O'Brien *et al*., 1997). They indicate a latest Arenig–earliest Llanvirn age for these strata and confirm that the contact between the middle and upper parts of the Exploits Group lies near the boundary separating Early and Middle Ordovician strata (Table 1; Figure 4).

Strata in the middle part of the Strong Island chert are intruded by the Llanvirn–Llandeilo Thwart Island gabbro (ca. 464 Ma) and are thus early–late Llanvirn in age (O'Brien *et al*., 1997). An upper age limit on the Strong Island chert is provided by the latest Llandeilo–earliest Caradoc black shales in the overlying Lawrence Harbour Formation. In the Lawrence Head–Upper Black Island area, the Strong Island chert represents a relatively thin but highly condensed Middle Ordovician succession.

In the Hummock Island limestone, the conodonts *Periodon aculeatus* and *Protopanderodus varicostatus* indicate a late Llanvirn–Llandeilo age; whereas, *Ansella jemtlandica, Polonodus tableheadensis* and *Periodon aculeatus* imply an early Llanvirn age (F.H.C. O'Brien, personal communication; O'Brien *et al*., 1997). The early Llanvirn fauna dates some of these bioclastic limestones as being older than most strata in the Cobbs Arm limestone (Fahraeus and Hunter, 1981). Significantly, the lower beds of the Hummock Island limestone also contain a reworked Arenig conodont population which include *Paracodylodus gracilis* and *Periodon flabellum*.

Plate 17. *Graded crystal-lithic wacke, located south of Michaels Harbour and west of Indian Arm Brook, is typical of the Strong Island chert of the Exploits Group. Note the preferred orientation of the voluminous green basalt lithoclasts as well as the sharp, planar, basal contact with grey chert.*

Plate 18. *Irregular zones of cream-coloured laminated chert and light grey, thin bedded siliceous siltstone, exposed east of Lewisporte Harbour, display quartz-rich nodules with alteration haloes. The mottled texture illustrated is characteristic of the upper part of the Strong Island chert.*

The absolute U/Pb zircon ages from the middle Tea Arm Formation and the Thwart Island gabbro, when combined with the biostratigraphic ages from the graptolite and conodont fauna in the New Bay Formation, the Lawrence Head Formation, the Strong Island chert and the Hummuck Island limestone, demonstrate that the Exploits Group (as defined herein) probably ranges in age from the Early Ordovician (earliest Arenig) to the Middle Ordovician (latest Llandeilo).

Correlation

The rhyolites and basalts in the lower Exploits Group are probably best correlated with similar volcanic rocks in the lower part of the Wild Bight Group (Swinden *et al*., 1990; O'Brien *et al*., 1997; MacLachlan, 1998). In particular, the graphitic and carbonaceous grey shale horizon in the Saunders Cove Formation of the Exploits Group is very similar to the dark grey shale horizon in the Omega Point Formation (Dean, 1977) of the Wild Bight Group. However, the New Bay Formation of the Exploits Group does not have a direct counterpart in the Wild Bight Group.

The Strong Island chert of the upper Exploits Group is probably the partial equivalent of the basal chert-dominated division of the Shoal Arm Formation and it may have correlatives in the uppermost Pennys Brook Formation of the Wild Bight Group (Table 1). Most of the Lawrence Head Formation and parts of the Hummock Island limestone of the upper Exploits Group may be equivalent to the upper part of the Summerford Formation and the Cobbs Arm limestone on New World Island (Kay and Williams, 1963).

Regional Interpretation

Island-arc tholeiites in the Little Arm East member had probably begun to accumulate in the earliest Arenig or latest Tremadoc. They were overlain by the high-silica sodic rhyolites of the Pushthrough member and directly succeeded by

Plate 19. *Margin-parallel banding in a vesicular equigranular sill of Thwart Island gabbro at Indian Cove in Ritters Arm is outlined by layered schlieren pegmatites. In the background, such gabbroic pegmatites are seen to occupy dilational transverse as well as longitudinal fractures.*

Plate 20. *Geopetal information (e.g., upper roof of magma chamber in direction of hammer head) is gleaned, in part, from the crescumulate texture of a Thwart Island gabbro sill near Point of Bay.*

the LREE-depleted pillowed basalts of the lower Pleasantview member. Swinden *et al*. (1990) interpreted such volcanic rocks to have crystallized from magmas whose source melts were generated in the lower crust of a rifting island arc and in the highly refractory parts of the underlying mantle wedge.

At this time, in the submarine volcanic environment of the Tea Arm arc, uplift of horst blocks probably controlled limestone deposition, diabase intrusion, chert veining, chlorite schist formation and hydrothermal stockwork mineralization. Subsequently, low-K calc-alkaline basalts and andesites erupted throughout most of the Pleasantview succession. However, in the uppermost part of this member, minor eruptions of boninitic basalt and refractory andesite heralded the development of the Saunders Cove iron formation. The Tea Arm and Saunders Cove formations of the Exploits Group have been interpreted as part of an Early Ordovician extensional primitive oceanic arc (O'Brien *et al*., 1997).

The sandy and silty turbidites of the Charles Brook member of the New Bay Formation are interpreted as turbidity current deposits formed on the lower parts of deep marine fans. The congomeratic turbidites of this member have intraclasts mixed with extrabasinal detritus and have been interpreted as filling fan channels (Helwig, 1967). Some large well-rounded volcanic clasts in conglomerate have geochemical signatures typical of flows in the Pleasantview member, suggesting that they may have been sourced in an uplifted and eroded part of the Tea Arm arc. In contrast, the Charles Brook lapilli stone horizons are considered to be pyroclastic avalanche deposits (Dec *et al*., 1993). The evidence of fragmentary graptolites, together with the predominance of mudstones occupying the Bouma E-interval, may indicate that the background sediments of the Brooks Harbour member had an outer fan slope or abyssal plain origin. The Tea Arm arc was apparently drowned in the Arenig during Brook Harbour time.

The transition from pelagic mudstone deposition in the Brook Harbour member to olistostrome formation in the

Saltwater Pond member was associated with a change in paleocurrent direction in the regional drainage pattern (Helwig, 1967). The presence of abundant detrital pillow basalt, the island arc geochemical affinity of large transported rafts of basaltic lava, and the similarity of the magmatic clasts in debrites to the tonalites and diorites of local igneous complexes suggests that the edifice skirted by the New Bay basin may have been part of a dissected volcanoplutonic arc (Hughes and O'Brien, 1994). At least parts of the Saltwater Pond member of the New Bay Formation may have originated as a slope apron deposit. In the middle part of the Saltwater Pond succession, strata were increasingly redistributed within the depositional basin by sliding on unstable Brooks Harbour muds. Hibbard (1976) considered rocks presently assigned to this unit to be one of the major protoliths of the blocks in the Dunnage Melange.

Plate 21. *Exposure on eastern Upper Black Island of the basal contact of the Hummock Island limestone (light-coloured strata) above the pillowed top of a basalt in the uppermost Lawrence Head Formation. The youngest conodonts present in these bioclastic limestone beds are earliest Llanvirn in age.*

The New Bay Formation has been interpretated as the arc-derived fill of a late Early Ordovician arc-rift basin (O'Brien *et al*., 1997). Although the remnant Tea Arm oceanic arc formed the substrate to this rift basin, the pyroclastic tephra in the Charles Brook member probably originated in a distal magmatically active arc, situated to the west (present coordinates) of the Wild Bight Group (cf. MacLachlan, 1998).

The eruption of the various within-plate basalts in the Lawrence Head Formation signalled a change in depositional environment, as the dynamic Exploits basin attempted to stabilize. Alkali basalts, similar to those in modern back arc basins, display the highest LREE enrichment and the highest concentrations of incompatible elements for any Exploits Group volcanic rock (O'Brien *et al*., 1997). Other Lawrence Head pillow basalts are slightly-enriched tholeiites with a characteristic REE pattern of positive Ta and negative Th and Ti anomalies. They are similar to modern basalts in mid-oceanic ridges or the spreading zones of marginal oceanic basins.

Fragments of dendroid graptolites are common in parts of the Strong Island chert (Williams *et al*., 1992). In living position, these benthic forms typically inhabited shallow marine waters in the near shore environment. Thus, the preservation of their fragile organic skeletons in a hemipelagic chert indicates relatively limited post-mortem transport from their original ecological setting. Moreover, the dendroids were rapidly buried in the deep-sea environment of the black siliceous argillites, the radiolarian-bearing ribbon cherts and the pelagic *U. austrodentatus* Zone and younger siltstones.

In some places, fragmentary dendroids have been recovered from grey Strong Island chert successions that contain thick interstratified sequences of strongly altered epiclastic turbidites. Hence, the variegated silicious mottled turbidite beds were probably altered in small depocentres flanked by shallow marine uplands.

The early Llanvirn conodont fauna of the Hummock Island limestone not only provides an independent younger limit for the cessation of back-arc volcanism in the Lawrence Head Formation but it gives the timing of the rise of the earliest Llanvirn volcanic rocks above carbonate compensation depths.The Llanvirn bioclastic carbonates lying directly above the Lawrence Head Formation also contain early–mid Arenig conodont species (O'Brien *et al*., 1997). The presence of two conodont populations in these bioclastic limestone beds may imply resedimentation of conodont-bearing Arenig carbonates contemporaneously with earliest Llanvirn upwelling of the back-arc lavas in the Exploits Group (O'Brien *et al*., 1997).

The Strong Island chert is interpreted as Middle Ordovician back-arc basin infill. The Hummock Island limestone is postulated to have covered a bathymetric high underlain by the Lawrence Head Formation, perhaps a volcanic plateau in the Exploits back arc basin.

Faulting during Accumulation of the Exploits Group

Whereas basalt, rhyolite and limestone of the Tea Arm Formation were buried beneath the lower Charles Brook member in certain parts of the New Bay arc-rift basin, they were apparently already eroded and uplifted in other areas prior to the deposition of the middle-upper portions of this member. A different part of the drowned Tea Arm arc became emergent in the late Arenig and provided detritus to the Saltwater Pond member.

Throughout the study area, a regionally extensive, relatively thin, locally condensed unit of Strong Island chert rests, in different localities, above the arc tholeiites of the Tea Arm Formation's Little Arm East member, the conglomeratic turbidites of the New Bay Formation's Saltwater Pond member and the alkali basalts at the top of the Lawrence Head Formation. Thus, the Strong Island chert straddles successions of primitive arc, arc-rift and back-arc strata in the Exploits Group. The intrabasinal topography buried by the deep-sea chert unit resulted from geological events other than the regional folding and thrust faulting of the Exploits Group.

Late Tremadoc–early Arenig pillow lavas come in direct contact with late Arenig–early Llanvirn pillow lavas where the Tea Arm and Lawrence Head formations are juxtaposed along the Paradise Fault (Figure 11). There, the mid–upper Tea Arm Formation, the Saunders Cove Formation and all of the New Bay Formation, which measure at least 2.5 km in stratigraphic thickness, are locally absent from the Exploits Group. This phenomenon is most simply explained by some combination of non-deposition, stratigraphic pinch-out, condensed stratigraphy, tectonic erosion, or structural excision of map units (Hughes and O'Brien, 1994).

The Paradise Fault may have been broadly coincident with a basin margin-fault that had its maximum growth as the New Bay Formation accumulated in the northeastern tract of the Exploits Group. Syndepositional movements on this structure may have begun with footwall uplift of the Tea Arm Formation and periodic emergence of Tea Arm pillow lavas above carbonate compensation depths. Subsequently, such faulting may have been the cause of the down-to-depocentre thickening of the New Bay Formation to the northeast of the Tea Arm footwall sequence. On the southwest side of the postulated growth fault, sediment starvation and local non-deposition of the upper Tea Arm, Saunders Cove and New Bay formations may have occurred near the uplifted fault block.

Wild Bight Group

The Wild Bight Group is located in the northwesternmost part of the Exploits Subzone examined for this report. Although a very small proportion of the total outcrop area of the Wild Bight Group is present in the study area, a wide variety of rock types and ages are present.

Distribution and Thickness

The Wild Bight Group is well exposed in the region between Glovers Harbour in Seal Bay and Osmonton Arm in New Bay (O'Brien, 1991a; MacLachlan, 1998; Figure 3). A poorly exposed tract of Wild Bight Group rocks occurs farther south between the communities of Point Leamington and Northern Arm (O'Brien, 1993a). The outcrop area of both tracts is approximately 30 km^2.

The total combined stratigraphic and structural thickness of the Wild Bight Group on the peninsula separating New Bay and Seal Bay is estimated to be in the order of 2 to 3 km (MacLachlan and O'Brien, 1998).

Stratigraphic Nomenclature

In 1977, P. L. Dean proposed what he considered to be the first regionally applicable stratigraphy for the entire Wild Bight Group. Listed in ascending order, these five formally defined lithostratigraphic units are the Omega Point Formation, the Sparrow Cove Formation, the Seal Bay Brook Formation, the Side Harbour Formation and the Pennys Brook Formation. This succession was held to be essentially continuously exposed about a regional fold termed the Seal Bay Anticline, which Dean (1977) positioned to the west of the present map area.

Swinden (1987) described eleven informal volcanic units within the Wild Bight Group in a study dealing mainly with chemostratigraphy. These partly overlapped with the original five formations but, as direct correlations were never made, the exact stratigraphic position of each of the eleven volcanic units was not reported. The Glovers Harbour, Long Pond, Nanny Bag Lake, Indian Cove and Side Harbour volcanic units were stated to contain depleted arc tholeiites and trondhjemitic rhyolites in association with massive sulphide deposits; whereas, the Seal Bay Bottom volcanic unit was generally considered to be dominated by calc-alkaline basalts and andesites. In contrast, the Northern Arm, Big Lewis Lake, Seal Bay Head, Badger Bay and New Bay volcanic units were thought to have a general absence of felsic volcanic rocks and to be typified by the occurrence of alkali basalts along with a subordinate amount of enriched calc-alkaline andesites. Swinden *et al*. (1990) proposed several different paleotectonic environments for volcanic strata that occur in the easternmost Wild Bight Group (e.g., the Glovers Harbour, Long Pond and Nanny Bag Lake units were petrochemically distinquished from the Northern Arm and Big Lewis Lake units).

MacLachlan (1998) did not formally name her subdivisions of the northeastern Wild Bight Group, although she stated that her "younger Wild Bight sequence" probably contained parts of the Omega Point Formation, the Seal Bay Brook Formation and the Pennys Brook Formation.

Lithostratigraphy

The stratigraphic base of the Wild Bight Group is nowhere exposed in the study area. However, the stratigraphic top of the group may be locally preserved at the contact with the Shoal Arm Formation in the Osmonton Arm area, in the Stowaway Pond–Bard Pond area, and in the area immediately north of New Bay Pond. There, volcaniclastic turbidites assigned to the Wild Bight Group pass stratigraphically upward into mottled grey, red, green or black chert successions, which have been previously assigned to a variety of map units in the Wild Bight Group and the Shoal Arm Formation.

The external boundaries of the Wild Bight Group are typically thrust faults, which affected rock units as young as the Late Ordovician Point Leamington Formation, and transcurrent faults, which affected rock units as young as the Botwood Group and the Hodges Hill batholith (Figure 3). The Northwest Arm Fault forms the eastern boundary of the Wild Bight Group and juxtaposes volcanosedimentary rocks in the upper part of the group with strata assigned to the Exploits Group and the Point Leamington Formation (Figure 11). The unit's southern boundary with the Botwood Group is the Northern Arm Fault and its northern boundary with the Shoal Arm Formation is a splay of the Lukes Arm Fault (Figure 3). Imbricate splays of the Northwest Arm Fault locally separate the Wild Bight Group from the South Lake Igneous Complex, the Lawrence Harbour Formation and the Shoal Arm Formation (Table 1).

Lithology

In the northern outcrop area of the Wild Bight Group, a variably thick sediment-dominated succession composed of green volcaniclastic wackes, conglomeratic turbidites, siliceous banded argillites and minor felsic pyroclastic rocks overlies calc-alkaline pillowed basalts associated with a bright-red cherty iron formation and a dark-grey laminated shale (MacLachlan, 1998). In the uppermost part of this generally coarse-grained and thick-bedded succession, thin interbeds of grey and red, siltstone turbidite, siliceous argillite and altered chert become dominant and attain a maximum 400 m thickness. The lower calc-alkaline basalts are well exposed from Cumlins Cove to Glovers Point along the eastern coast of the Main Channel south of Leading Tickles; whereas, the upper sedimentary strata outcrop extensively between the Thimble Tickles area in the west and the Osmonton Arm area in the east. Interstratified lenticles of subalkaline to alkaline basalt flows are poorly developed in the upper turbidite sequence west of Osmonton Arm (MacLachlan, 1998).

In the southern outcrop area of the Wild Bight Group examined for this report, calc-alkaline and alkaline basalts are present as mappable lenticles in the volcaniclastic turbidite sequence overlying the bright-red cherty iron formation and a dark-grey laminated shale (O'Brien, 1993a; McConnell and O'Brien, 2000). Here, the maximum stratigraphic thickness of the upper volcanosedimentary sequence of the Wild Bight Group is approximately 1.5 to 2 km thick.

The chert and wacke successions that interdigitate with calc-alkaline or alkaline pillow lavas in the eastern Wild Bight Group, especially those in the northeast of the study area, are relatively thin in comparison with those in the western Wild Bight Group (O'Brien, 1997; McConnell and O'Brien, 2000). They are best correlated with the Pennys Brook Formation, the stratigraphically highest and thickest formation of the Wild Bight Group.

Many of the largest mafic intrusions in the eastern Wild Bight Group are gabbro sills hosted by the Pennys Brook Formation. In Osmonton Arm, they are observed to display peperitic margins with thinly bedded Pennys Brook turbidites. Thus, they are probably coeval with the Gummy Brook gabbro sills in the western Wild Bight Group (O'Brien and MacDonald, 1996, 1997) and the Middle Ordovician Thwart Island gabbro sills in the eastern Exploits Group (O'Brien *et al*., 1997).

A relatively fresh alkali gabbro crosscuts highly altered rocks of the Glovers Harbour volcanic unit near the Lockport massive sulphide prospect (Figure 14). Diabase dykes intrude this Early Ordovician gabbro body and are also widespread in the Glovers Harbour, Long Pond and Nanny Bag Lake volcanic units of the eastern Wild Bight Group (Plate 22). Mafic dykes are commonly seen in swarms in the vicinity of the stockwork alteration zones observed within these units.

Age

Thin beds of graptolite-bearing black shale and coral-bearing limestone conglomerate are locally present in a thinly bedded, light grey, siliceous laminated argillite and siltstone turbidite succession in the uppermost part of the Pennys Brook Formation. Biostratigraphically useful graptolites, including *Nemagraptus gracilis* itself, were collected in Osmonton Arm southwest of Mussel Bed Island. They indicate a late Llandeilo–early Caradoc age for this youngest dated unit in the Wild Bight Group (S.H. Williams, unpublished data; 1992; *see N. gracilis* Zone on Table 1). A carbonate conglomerate located near the Mill Pond trail to Cramp Crazy Lake occurs lower in the Pennys Brook Formation; however, coralline limestone clasts in this conglomerate proved to be barren of conodonts (F.H.C. O'Brien, personal communication, 1993).

The Glovers Harbour volcanic unit, which was originally grouped with Dean's (1977) Pennys Brook Formation, is now known to be early Arenig or older in age on the basis of an intrusive contact with a gabbro sill, isotopically dated at ca. 486 Ma (G.R. Dunning, unpublished data, 1994; MacLachlan and Dunning, 1998a).

Volcanic and Sedimentary Lithodemes

Internal lithostratigraphic subdivision of the Wild Bight Group has been hampered, not only by complex strata- parallel faults, but also by primary lithofacies variations and a general lack of biostratigraphic control. Nevertheless, it appears that most of the peninsula separating Seal Bay and

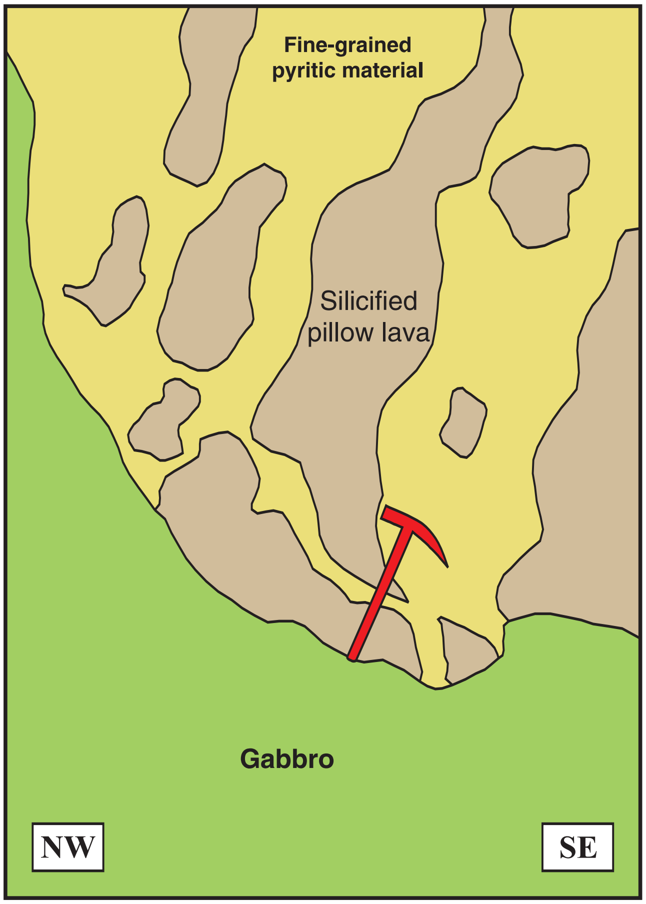


Figure 14. *Sketch drawn from photograph taken near the Lockport prospect looking northeastward at vertical cliff section [E609450 N5480000] of a gabbro dyke (bottom of drawing) and steeply inclined horizons of pillow breccia (top of drawing). Note relict pillow structure in these pyritized and silicified basalts. Note also the crosscutting intrusive contact of the unaltered gabbro, which is dipping gently northeastward in the foreground near the geological hammer.*

New Bay is underlain by cyclic turbidite lithodemes typical of the Pennys Brook Formation, the type area of which is located some 20 km farther west in the Wild Bight Group (Dean, 1977; O'Brien, 1997). However, east and west of Glovers Harbour, smaller-size map units dominated by red chert and interbedded olistostrome and by green siltstone and interbedded pillow lava (MacLachlan, 1999) may possibly represent parts of the Omega Point and Sparrow Cove formations, respectively.

Kate MacLachlan was the first worker in the Wild Bight Group to demonstrate an early Middle Ordovician age for volcanosedimentary lithodemes thought to lie below or near the base of the Pennys Brook Formation (MacLachlan, 1998; MacLachlan and Dunning, 1998b). As a result of her work, these lithodemes were assigned to a position stratigraphically above the Glovers Harbour volcanic unit and stratigraphically below the *N. gracilis* Zone argillites and siltstone turbidites of the upper Pennys Brook Formation.

Rhyolite–Basalt Lithodemes

Flow banded rhyolites and pillowed basalts of the Glovers Harbour volcanic unit and the Long Pond volcanic unit were grouped in MacLachlan's (1998) "lower Wild Bight sequence". In several localities between Leading Tickles and Northern Arm, they lie tectonically adjacent to the tonalites and diorites of the Ordovician South Lake Igneous Complex (MacLachlan, 1999). On the basis of existing geochronology and geochemistry, MacLachlan and Dunning (1998a) argued that these intrusive and extrusive suites not only formed at the same time in the early Arenig but that they were geochemically and tectonically related to each other. In many areas, these bimodal volcanic and plutonic map units are directly juxtaposed against the sedimentary and within-plate volcanic successions found in the Middle Ordovician part of MacLachlan's (1998) "upper Wild Bight sequence".

Red Chert–Olistostrome–Basalt Lithodemes

Detailed mapping in the Thimble Tickles area (O'Brien, 1991a) indicated that a thick Pennys Brook unit of tuffaceous conglomerate, ferruginous wacke and green siliceous argillite underlay a more extensive but thinner Pennys Brook unit of grey-green ribbon chert and red parallel-laminated argillite. There, the lithodemes characteristic of the lower of the two Pennys Brook sequences pinch out to the southwest above a substrate of pillow lavas, which MacLachlan (1998) showed were calc-alkaline and similar to those in the Sparrow Cove Formation (Swinden *et al.*, 1990). On Ward Island and islands farther south, the sedimentary lithodemes above the calc-alkaline pillow lavas include spectacular debris flows rich in basalt clasts, polymict conglomerates having well-rounded boulders of granite and outsized blocks of exotic limestone, and sand-matrix olistostromes containing slumped fragments of red chert and basalt breccia.

In the northeastern Wild Bight Group, the Glovers Harbour volcanic unit is separated from the Shoal Arm Formation by less than 1500 m of volcanosedimentary strata. Yet, as a whole, the Wild Bight Group probably has a similar age range and is up to 4 km thick.

Basalt–Wacke–Argillite Lithodemes

Thin-bedded, fine-grained, shard-rich felsic tuffs, which are not associated with any known mineralization, occur above calc-alkaline pillow lavas and graded pillow breccias in the Glovers Point and Cumlins Head areas in the northeastern Wild Bight Group. They lie below pebbly to sandy wackes and banded siliceous argillites in the Leading Tickles and Mill Pond areas. MacLachlan and Dunning (1998b) dated these and other felsic tuffs and originally established their late Arenig–early Llanvirn depositional age (ca. 473 Ma to 470 Ma).

This early Middle Ordovician volcanic and sedimentary sequence underlies and is intertongued with the Long Island volcanic unit (the within-plate pillow lavas of Osmonton Arm). Moreover, the alkali basalts of Swinden's (1987) Seal Bay Head volcanic unit appear to be contained within this same sequence of Pennys Brook lithodemes in the Wild Bight succession near Inner Seal Head.

In the southeastern Wild Bight Group, calc-alkaline basalts of the Northern Arm volcanic unit display enriched LREE profiles with positive Th and La and negative Nb anomalies (McConnell and O'Brien, 2000) and crop out, in places, at the faulted contact with the Shoal Arm Formation (Figures 3 and 15). However, in other localities, within-plate basalts interstratified with the Pennys Brook Formation (e.g., New Bay and Long Island volcanic units) occupy a similar structural position.

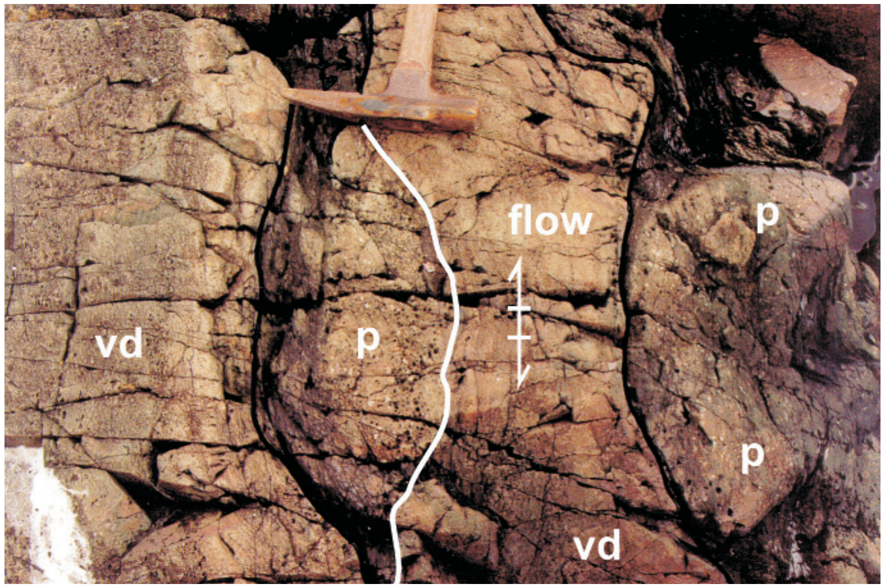


Plate 22. *Plan view of subvertical dykes of vesicular diorite (vd), which display chilled margins (intrusion at left) and vertical flow banding (central intrusion). Located in a mafic dyke swarm within the Glovers Harbour volcanics at Thimble Tickle bottom, the Wild Bight country rocks are variably foliated pillow lavas (p) that have their vertical enveloping surface striking parallel to the hammer head.*

A detailed illustration and comprehensive discussion of the lithodemic associations and informal internal units in the eastern Wild Bight Group is reported in MacLachlan (1998), MacLachlan and Dunning (1998a; 1998b), and MacLachlan and O'Brien (1998).

Correlation

MacLachlan (1998) correlated the bimodal basalt–rhyolite association in Swinden's (1987) Glovers Harbour volcanic unit of the Wild Bight Group with the early Arenig bimodal basalt–rhyolite association in the Tea Arm Formation of the Exploits Group (O'Brien *et al*., 1997). The Long Pond volcanic unit and the Nanny Bag Lake volcanic unit of the southeastern Wild Bight Group (Swinden, 1987) were correlated with the Glovers Harbour volcanic unit (MacLachlan and O'Brien, 1998).

Swinden and Jenner (1992) included the New Bay Pond volcanic unit (Swinden, 1987) in the Wild Bight Group. Accordingly, the Silurian Frozen Ocean Group of Dean (1977) has been informally abandoned (Figure 3). The New Bay Pond arc tholeiites are interstratified with mineralized felsic volcanic rocks that are geochemically similar to those in the correlative Glovers Harbour volcanic unit (Swinden, 1988).

MacLachlan (1998) demonstrated that the early Arenig or older rhyolites and basalts of the Glovers Harbour volcanic unit belonged to her "older Wild Bight sequence" and that they were previously incorrectly assigned to the Pennys Brook Formation (Dean, 1977). She also interpreted an early Llanvirn sequence of calc-alkaline basalts in the northeastern part of the Wild Bight Group (part of her "younger Wild Bight sequence") to lie stratigraphically within and above a local marine redbed unit correlative with the Omega Point Formation. She considered these Middle Ordovician (post-Omega Point) volcanic units to be completely absent in the easterly adjacent Exploits Group. However, she postulated that tectonic fragments of this mafic to intermediate volcanic arc might be found farther west in the upper Victoria Lake Group and possibly the Roberts Arm Group (Table 1).

The uppermost Wild Bight Group contains lithological correlatives of the Llanvirn–Llandeilo Strong Island chert of the Exploits Group (O'Brien, 1990; Williams *et al*., 1992). The unit is well exposed in the outer reaches of Osmonton Arm, on Long Island in Osmonton Arm and on Alcock Island off Leading Tickles.

Regional Interpretations

Swinden (1987) and Swinden *et al*. (1990) interpreted the overall chemostratigraphic record of the Wild Bight Group as indicating a gradual transition from the environment of an intraoceanic volcanic island arc to a magmatically actively intra-arc rift zone and, with further extension, to the environment of an ensimatic back-arc basin.

MacLachlan (1998) concurred that the Early Ordovician LREE-depleted arc tholeiites and high-silica sodic rhyolites in Swinden's (1987) Glovers Harbour volcanic unit constituted a primitive extensional oceanic arc sequence but

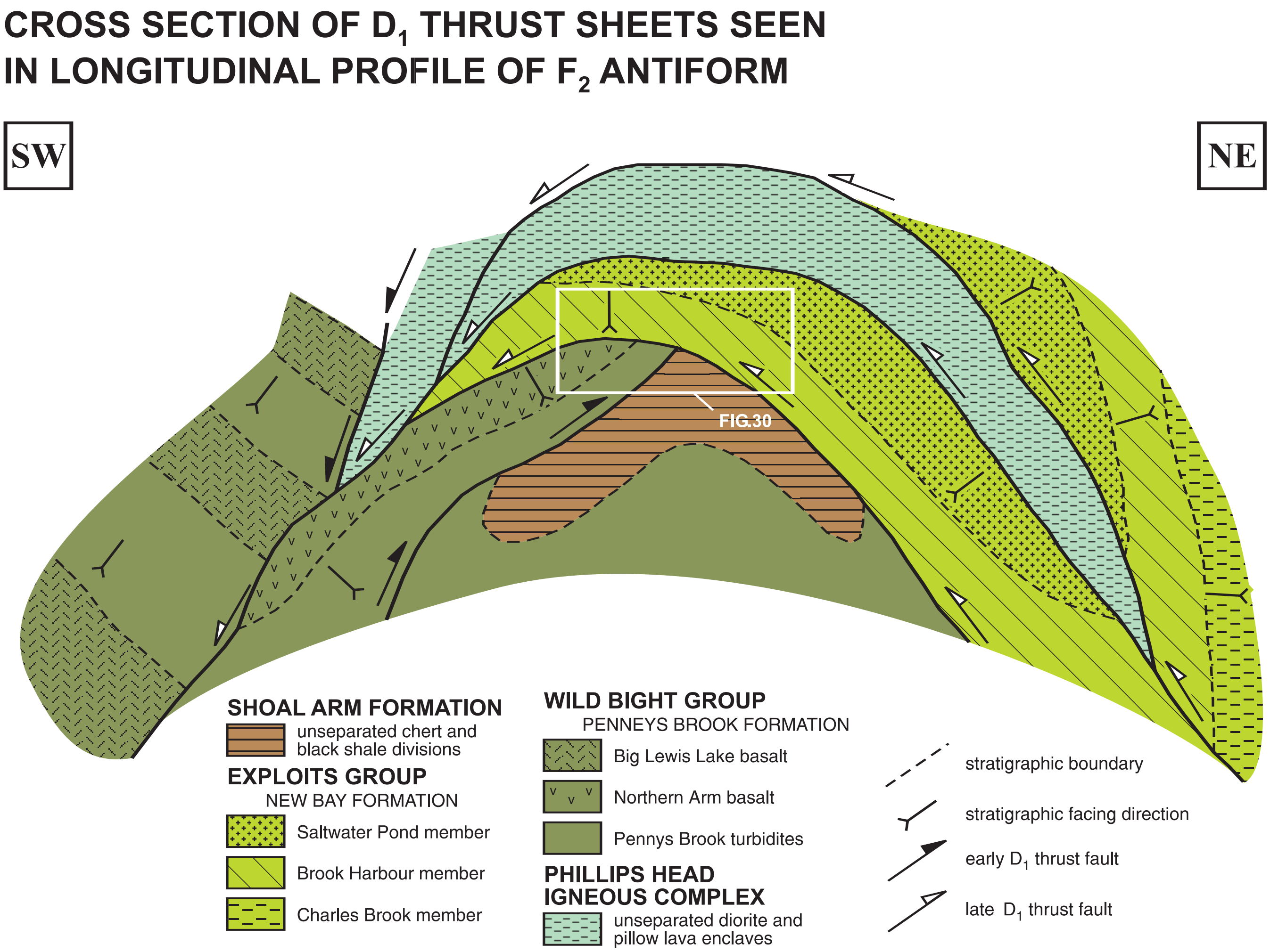


Figure 15. *Unbalanced cross section of the Phillips Head igneous complex and adjacent stratified rock units, drawn immediately northwest of the Northern Arm Fault. Section viewed looking northwestward, up-the-plunge of a D_1 antiformal culmination within a post-Shoal Arm Formation thrust stack.*

thought that the bimodal unit of her "older Wild Bight sequence" had originally laid beneath the red iron formation of Dean's (1977) Omega Point Formation.

Where exposed in the Thimble Tickles area, the boundary between the "upper Wild Bight sequence" wedge of Middle Ordovician strata and the Early Ordovician basalts and intercalated red cherts of the Glovers Harbour volcanic unit is strongly modified by a regional ductile shear zone. However, MacLachlan (1998) believed that the provenance of magmatic clasts in olistostrome and conglomerate horizons in the Middle Ordovician succession indicated that plutonic and volcanic rocks in units such as the South Lake Igneous Complex and the Glovers Harbour volcanic unit were already together in what was the source area of the northeastern Wild Bight Group by the early Llanvirn.

MacLachlan (*op. cit.*) interpreted the early Middle Ordovician volcanic and sedimentary strata in the eastern Wild Bight Group to represent the infill of a rift basin within a mature continental margin volcanic arc, which developed above the older "Glovers Harbour" remnant oceanic arc.

Lawrence Harbour Formation, Shoal Arm Formation and Luscombe Formation

Black graptolitic shale comprises parts of several well-exposed Middle–Late Ordovician lithostratigraphic units in the Notre Dame Bay region of central Newfoundland (Figure 3). These formations are the oldest, most widespread and diagnostic units within the overstep sequence of the Exploits Subzone (e.g., Williams, 1991; Williams *et al*., 1992). Middle–Late Ordovician black shales comprise thin marker-bed sequences within several major rock groups in central Newfoundland, although they are commonly unnamed or informally defined (Williams, 1993; Table 1).

Distribution and Thickness

In the western part of the study area, the Shoal Arm and Lawrence Harbour formations crop out on the fault-modified flanks of regional anticlinoria, lying above the Early–Middle Ordovician Wild Bight and Exploits groups, respectively. These older rock groups occur on opposing limbs of fault-bounded synclinoria that are cored by black shales and sandstone turbidites of proven early Caradoc to mid Ashgill age. Farther east, where the Luscombe Formation lies adjacent to the Dunnage Melange, the overlying late Caradoc to mid Ashgill succession is locally condensed, and fault-bounded synclinoria flanking the black shales contain turbidites as young as the mid Llandovery (Table 1).

Since the Lawrence Harbour, Shoal Arm and Luscombe formations have many internal faults, estimates of map unit thickness represent the total tectonic thickness of each formation. The Lawrence Harbour Formation varies from as little as 250 m thick in its type area to as much as 600 m thick in Osmonton Arm bottom. The variable amounts of black graptolitic shale present in this formation are generally reflected in the thickness estimates. In contrast, the Shoal Arm Formation contains relatively little black shale and is mostly composed of interbedded chert and wacke. It is generally thicker than the Lawrence Harbour Formation but shows less thickness variation along strike. In the Shoal Arm area, this formation is about 500 m thick; whereas, it is approximately 800 m thick in the New Bay Pond–Big Lewis Lake area. The Luscombe Formation is an approximately 350 to 400 m thick unit, although green wacke and variegated chert predominate over dark-grey argillite and black shale.

Stratigraphic Nomenclature

In 1969, J.A. Helwig defined his basal Lawrence Harbour Shale to rest above the pillow lavas of the Lawrence Head Formation of the Exploits Group. He also defined the top of an unnamed division of Caradoc black argillite to lie below sandstone turbidites now assigned to the Point Leamington Formation of the Badger Group (Williams, 1991). Because the red and green strata of the Strong Island chert and the overlying grey bioturbated chert are only about 10 m thick in Lawrence Harbour bottom, this was a convenient way to locally define the base of the Lawrence Harbour Shale and to distinguish these black shales and cherts from younger parts of the black shale formation.

The western Bay of Exploits region contains the type area and reference section of the Lawrence Harbour Formation (Williams, 1995; Figure 3). As proposed by Williams (1995), the bottom of the formation is defined by the lowest black graptolitic shale above grey bioturbated chert. He drew the top of the Lawrence Harbour Formation at the lowest sandstone turbidite above black graptolitic shale. Thus, correctly or not, most of the grey bioturbated chert (i.e., the regionally extensive'lower' Lawrence Harbour Shale) has been removed from the Lawrence Harbour Formation.

The Shoal Arm Formation was originally recognized and defined by Espenshade (1937) in the Badger Bay area, which occurs to the northwest of the region discussed in this report (Figure 3). He grouped the distinctive black shales and multicoloured cherts of the Shoal Arm Formation with the massive granular wacke of the stratigraphically overlying Gull Island Formation and several other units in the Badger Bay Series (Table 1).

Dean (1977) extended a revised version of Espenshade's Shoal Arm Formation into the New Bay Pond–Northern Arm area (Figure 3), although he referred to the overlying massive wackes as the Sansom greywacke (Table 1). In the New Bay Pond area, Kusky (1985) suggested that the Shoal Arm Formation be elevated to group status, and informally erected and named what he considered to be its three constituent formations.

The Luscombe Formation was originally defined as a lithostratigraphic unit by Kay (1975), who included it within his Campbellton sequence.

Lithostratigraphy of the Lawrence Harbour Formation

Despite having conformable lower and upper stratigraphic boundaries, it is practical to exclude the Lawrence Harbour Formation and succeeding units from the Exploits Group (Helwig, 1969), as its biostratigraphic or lithostratigraphic equivalents occur in several other rock groups in central Newfoundland (Kean *et al*., 1981).

In the New Bay area, the stratigraphic top of the Lawrence Harbour Formation is positioned at its gradational contact with thinly bedded sandstone turbidite. Less commonly, it is placed at the sharp erosive contact with olistostrome and overlying thickly bedded pebbly wacke. The stratigraphic base of the Lawrence Harbour Formation is placed at its gradational contact with the Strong Island chert or its abrupt irregular contact with the Hummock Island limestone.

Lithology

The Lawrence Harbour Formation has several distinctive subunits which are developed over a large part of central Notre Dame Bay. However, as each subunit is relatively thin, they cannot be individually surveyed on 1:50 000 scale geological maps.

The informal constituents of the Lawrence Harbour Formation (O'Brien, 1990) are a lower subunit composed of a dark-grey laminated chert, a thinly bedded, graded, mottled-textured siltstone, and a light-grey bioturbated chert having conspicuous black siliceous argillite laminae. It is succeeded by an intermediate subunit distinguished by laminated or very thinly bedded, black pyritiferous siltstones which contain recessed partings of black graptolitic shale. A highly fossilifereous upper subunit is made up of a black carbonaceous shale. It has an increasing proportion of interlaminated grey shale ascending the upper portion of the Lawrence Harbour Formation.

In the past, most workers in the study area have regarded a grey bioturbated chert subunit as the basal division of the Lawrence Harbour Formation.

Age

In the study area, the black shales of the Lawrence Harbour Formation have yielded graptolites indicative of the *Nemagraptus gracilis* Zone, the *Climactograptus bicornis* Zone and the *Dicranograptus clingani* Zone (Williams and O'Brien, 1994; Williams, 1995). Thus, the Lawrence Harbour Formation ranges in age from the latest Llandeilo–early Caradoc, through the mid Caradoc and into the latest Caradoc (Table 1). Although the lowest subunit of mottled and bioturbated chert is apparently devoid of macrofossils, it occurs conformably below the first continuous bed of *N. gracilis* Zone black shale.

Correlation

The black graptolitic shales of the Lawrence Harbour Formation are probably equivalent to those in the Shoal Arm Formation in western Notre Dame Bay and the Luscombe, Rogers Cove and Dark Hole formations in eastern Notre Dame Bay (Table 1).

Lithostratigraphy of the Shoal Arm Formation

A distinctive red chert of the Shoal Arm Formation is observed to stratigraphically overlie the Pennys Brook Formation of the Wild Bight Group in the type area in Badger Bay and also in the New Bay Pond area. It has been reported to contain poorly preserved radiolaria (Kusky, 1985; Bruchert *et al*., 1994). The upper boundary of the Shoal Arm Formation is observed where black graptolitic shales underlie unfossiliferous, thick-bedded to massive, granular wackes. These proximal wackes are lithologically similar to beds in the type area of the Gull Island Formation of the Badger Group and they are notably devoid of the graptolitic siltstone intervals typical of the Point Leamington Formation [*see* Dickson (2000) for alternative nomenclature]. In places, a tectonically sheared, graptolite-bearing sand-matrix olistostrome occurs at the top of the Shoal Arm black shale sequence. In the past, various workers have included it within the Shoal Arm Formation or excluded it from of that formation.

Lithology

Most workers consider the Shoal Arm Formation to be composed of a basal subunit of red radiolarian chert with minor turquoise chert and grey manganiferous siltstone turbidite, a gradationally overlying intermediate subunit of distinctively mottled sandstone turbidite, interstratified red and green siliceous argillite, and grey bioturbated chert, and a conformably overlying uppermost subunit dominated by black carbonaceous shale. The limited thickness of each of these lithological constituents of the Shoal Arm Formation prevents the highly distinctive subunits from being individually surveyed on 1:50 000 scale geological maps.

Age

The uppermost part of the grey bioturbated chert in the Shoal Arm Formation lies within the *C. bicornis* Zone and the overlying black shales are known to span parts of the *C. bicornis* Zone and the *D. clingani* Zone (Table 1). However, the lowest Badger Group sandstones locally occur in the *D. clingani* Zone and thus they are also Caradoc. Slump-folded black shale olistoliths in the olistostrome at the boundary between the Shoal Arm and Gull Island formations have yielded graptolites indicative of the *D. clingani* Zone (Williams, 1993; Williams and O'Brien, 1994).

Correlation

The lower Shoal Arm Formation may be correlated, in part, with the lower Luscombe Formation. Subunits from both formations may be partial equivalents of the Llanvirn–Llandeilo Strong Island chert.

Lithostratigraphy of the Luscombe Formation

The stratigraphical base of the Luscombe Formation is drawn where thinly bedded cherts and siliceous laminated argillites overlie the pillowed basalt, pillow breccia and mafic tuff of the Lawrence Head Formation of the Exploits Group (formerly the Loon Harbour volcanics of the Campbellton sequence). The stratigraphical top of the formation occurs above graptolitic black shales at the lowest bed of the overlying sandstone turbidite succession.

The Luscombe Formation is directly overlain by a variable thickness of green–grey mottled sandstone turbidites, siliceous black laminites, graded siltstone rhythmites and silicified pebbly wackes, which together comprise the basal part of the Campbellton greywacke (formerly the Riding Island greywacke of the Campbellton sequence). These are succeeded by the thick-bedded, graded, granular wacke and thinner bedded sandstone turbidites typical of parts of each of the Gull Island, Point Leamington, Campbellton and Sansom greywacke units.

Graptolitic black shales lying above limestone and bioturbated chert on Upper Black Island range into the uppermost Caradoc (S.H. Williams, 1993; personal comunication, 1993). As in the North Campbellton and Loon Harbour areas, such strata lie adjacent to faulted tracts of the Exploits Group and the Dunnage Melange. If the black shales of northeastern Upper Black Island represent the upper part of the Luscombe Formation (Figure 3), then the upper stratigraphic contact of this unit is also locally defined by an early Ashgill olistostrome (*see* Badger Group, page 46).

Lithology

The Luscombe Formation is mainly composed of interstratified beds of grey laminated chert, grey-green tuffaceous wacke and grey siliceous argillite. It contains lesser amounts of iron-rich turquoise chert, mottled silicified wacke, manganese-rich dark grey chert, green siliceous argillite, carbonaceous siltstone and black shale.

Many of the grey cherts in the Luscombe Formation situated below the graptolitic black shale and the iron oxide-weathered, dark-grey siltstone and laminated argillite are amorphous. Other grey cherts illustrate textures related to the replacement of relict primary bedforms. These cherts and associated siliceous argillites have been reported to contain manganese carbonate, garnet, pyrolusite, pyrite and pyrrhotite in different levels of the Luscombe succession (Sangster, 1993).

Age

Graptolites from the Luscombe Formation indicate that the stratigraphically highest subunit of black shale belongs to the *D. clingani* Zone (upper Caradoc) and that at least some of immediately underlying cherts and wackes in the Luscombe Formation are Caradoc in age (Table 1). Faunas representative of the *D. clingani* Zone were collected near the Old Iron Mine on the north limb of the Campbellton Syncline and near the bridge over Indian Arm Brook on the south limb of the syncline (Williams, 1993).

Correlation

Some thin beds of shard-rich tuff and thick beds of basalt-rich wacke near the base of the Luscombe Formation are similar to those in the transition zone between the Strong Island chert and the Lawrence Head Formation of the Exploits Group.

The Llandeilo–Caradoc Interval in Central Notre Dame Bay

New data collected during this study on the distribution of graptolite biozones has provided constraints on the stratigraphic evolution of the Llandeilo–Caradoc formations of central Notre Dame Bay. Black shales appear to have been deposited for various lengths of time in different units in separate parts of the region (Williams and O'Brien, 1994).

The maximum time interval of shale deposition is represented by the Lawrence Harbour Formation (*N. gracilis* Zone, *C. bicornis* Zone and *D. clingani* Zone), a lesser duration of time within the Shoal Arm Formation (*C. bicornis* Zone and *D. clingani* Zone) and the minimum interval in the Luscombe Formation (*D. clingani* Zone). In the Lawrence Harbour Formation, Caradoc cherts are restricted to the thin basal subunit and presumably spanned only a part of the *N. gracilis* Zone. Deposition of chert lasted throughout a larger interval of the Caradoc in the Shoal Arm Formation, as the grey bioturbated chert division resides in the *C. bicornis* Zone and the *N. gracilis* Zone lies, at least locally, in the upper Pennys Brook Formation of the Wild Bight Group. Chert formation was only briefly interrupted by deposition of *D. clingani* Zone black shale within the succession ranging from the lower Luscombe Formation to the lower Campbellton greywacke.

With regard to the accumulation of younger siliciclastic turbidite deposits, those above the Shoal Arm black shale began earlier than some, though not all, of those above the Lawrence Harbour black shale. Early onset of polymict conglomerate deposition above the Shoal Arm and Lawrence Harbour formations occurred where olistostromes are locally scoured into *D. clingani* Zone shales. In the *D. clingani* Zone of the Shoal Arm Formation, accumulation of sandstone turbidites had begun before olistostrome emplacement. In contrast, prior to the same debris flow-forming event, Lawrence Harbour strata only record evidence of quiescent pelagic deposition.

The ages of graptolites from the different fossil collections made in the area surveyed verify the lithostratigraphic order of the Lawrence Harbour, Shoal Arm and Luscombe formations, independently document their internal tectonic duplication, and establish their biostratigraphic affinity with other Caradoc black shale–chert formations in the Exploits Subzone (Williams and O'Brien, 1994; Hughes and O'Brien, 1994).

Regional Interpretations

The widespread deposition of Caradoc pelagic shale marks the drowning of Cambro-Ordovician magmatic arc, arc-rift and back-arc complexes during a high-stand of Iapetan sea level. It also heralds the cessation of deep-marine volcanism in the Exploits Subzone. Historically, some of these pelagic shale-bearing formations were postulated to represent Middle Ordovician ocean floor muds that were encrusted with manganese nodules near deep-water vents (Kay, 1975).

Although dominated by anoxic deposits, the Caradoc shale–chert formations within the central Notre Dame Bay region display internal subunits of disparate lithology and thickness, subtly different biostratigraphical age ranges, and external (lithostratigraphic) boundaries that indicate along-strike variations within and between map units. The development of nodular, sulphidized, manganese or ferroan carbonate-rich beds is related to localized alteration and replacement of siliciclastic strata. Such rocks are found within certain parts of the Luscombe and Shoal Arm formations and in the sedimentary strata lying beneath these formations. The preferred stratigraphic setting seems to be in relatively thick, early Llanvirn–early Caradoc chert-dominated successions capped by notably young or relatively thin intervals of black graptolitic shale.

Tectonic Implications

Black graptolitic shales ranging from the mid to upper Caradoc are located in the upper part of the Lawrence Harbour, Shoal Arm and Luscombe formations. Although their tectonic thickness is generally in the order of 200 to 300 m, strata from the *C. bicornis* and *D. clingani* biozones commonly occur in a lithotectonic sequence which is not in its original biostratigraphic order. As map unit separation across thrust faults is considered in terms of the stratigraphic thickness omitted (and not in terms of the missing time intervals in a lithotectonic sequence), minor displacements in a condensed fault-imbricated succession commonly result in large age gaps between the hanging wall and footwall plates. Consequently, the throw and heave of fault structures at the external boundaries of Caradoc black shale formations have generally been overestimated in central Notre Dame Bay.

Badger Group

The term Badger Group is useful to describe, as a single tectonic entity, all marine sedimentary strata in the overstep sequence of the Exploits Subzone west of the Dog Bay Line (Williams *et al*., 1993, Williams, 1995d) that formed before the Silurian onset of terrestrial volcanism and after the latest Middle to Late Ordovician drowning of Dunnage Zone rocks and their burial by pelagic black shale. In the area surveyed, the Badger Group is represented by the following units, named as follows: the Gull Island Formation, the Point Leamington greywacke, the Point Leamington Formation, the Sansom greywacke, the Sansom Formation, the Campbellton greywacke, the Riding Island greywacke, the Upper Black Island greywacke, the Goldson conglomerate, the Goldson Formation, the Lewisporte conglomerate, and the Randels Cove conglomerate (Figure 3; Table 1).

In central Notre Dame Bay, the formations and informal units of the Badger Group are distinguished by different associations of siliciclastic and clastic carbonate rocks and by various biostratigraphic ranges for these lithofacies. They also illustrate variations in their total thickness or in the thickness of individual subunits, variations in the grain size and degree of stratification on the scale of an entire unit, and variations in the coarse clastic or fine hemipelagic component of certain units. Badger Group units show important changes in the total representation of olistostromes or broken formations in any one section and also illustrate changes in the internal stratigraphic positions of such olistostromes.

Distribution and Combined Thickness

In the study area, the Badger Group is the most widely developed succession within the overstep sequence of the Exploits Subzone. The most extensive tract of contiguous strata in the Badger Group in central Notre Dame Bay is well exposed along the western fjords of New Bay, where the group crops out over a 100 km^2 area. A combined structural and stratigraphic thickness of approximately 3 km is estimated for this particular fault-bounded tract of the Badger Group.

The Badger Group is exposed in different regional fault blocks in the area surveyed. In the west, where the Badger Group overlies the Exploits and Wild Bight groups, Williams (1995d) referred to the area as the Badger Belt. In the east, where the Badger Group underlies the Botwood Group, Williams (*op. cit.*) referred to the area as the Botwood Belt. However, regarding the characteristics of map units in the Badger Group, there is as much variation within a belt as there is between belts (*see below*).

Stratigraphic Nomenclature

Some lithological units presently assigned to the Badger Group were previously given a formal stratigraphic rank in geographically restricted areas of Notre Dame Bay. Earlier workers informally named other map units in the Badger Group on the basis of a correlation with the type area of a particular formation or group. An informal designation is used where the intervening ground has not been systematically mapped on 1:50 000 scale by earlier workers.

In central and eastern Notre Dame Bay, the Gull Island Formation, the Point Leamington Formation, the Sansom Formation and the Goldson Formation are formally defined lithostratigraphic units. The Point Leamington greywacke, the Sansom greywacke, the Campbellton greywacke, the Riding Island greywacke, the Upper Black Island greywacke, the Goldson conglomerate, the Lewisporte conglomerate and the Randels Cove conglomerate are mappable informal units. The Campbellton greywacke, the Upper Black Island greywacke, the Lewisporte conglomerate and the Randels Cove conglomerate are new terms used in this report.

Lithostratigraphy

The Badger Group of central Newfoundland comprises a regionally extensive, calcareous and quartz-rich turbidite succession which has abundant olistostrome horizons and soft-sediment deformation features indicative of deposition in tectonically active basins (Pickering, 1987). However, it does not contain intercalated volcanic strata.

In most places in the northern part of central Newfoundland, the lower stratigraphic boundaries of Badger Group formations are with black or grey shales of proven or purported latest Caradoc or earliest Ashgill age (*see* previous section). In one locality on New World Island, the Badger Group has been reported to unconformably overlie pre-Caradoc volcanic strata assigned to the Summerford Formation (Williams, 1995e; Figure 3).

The upper stratigraphic boundaries of Badger Group formations are not observed in the area surveyed. However, Badger Group strata are also reported from eastern Notre Dame Bay, where they are purported to conformably underlie the lowest terrestrial basalt formation of the Botwood Group (Currie, 1993; Williams *et al*., 1993). Farther east in the regional overstep sequence, Badger Group turbidites lie tectonically adjacent to shallow marine rocks of the Indian Islands Group, which is unconformable on the Hamilton Sound Group (Figure 3; Table 1). The upper part of the Indian Islands Group is considerably younger than the terrestrial strata of the Botwood Group and the underlying Badger Group (Boyce *et al*., 1993; 1994).

Age

In the study area, fossil-bearing strata assigned to the Badger Group have an age range which is known to extend from the mid Late Ordovician (latest Caradoc–earliest Ashgill) to the late Early Silurian (latest Llandovery). In this report, the biostratigraphic range of each lithostratigraphic unit mapped in central Notre Dame Bay is presented within the individual description of that unit. It is also portrayed in Table 1.

Correlation

Certain formations of the Badger Group may have partial equivalents in other lithostratigraphic units of the Newfoundland Dunnage Zone. In the Exploits Subzone, such strata may not have everywhere been separated from the Davidsville and Bay d'Espoir groups. Similarly, in the Notre Dame Subzone, such strata may reside, correctly or not, in the Roberts Arm Group (Crescent Lake Formation) and the upper part of the Windsor Point Group.

Point Leamington Formation, the Goldson Conglomerate and the Randels Cove Conglomerate

The Point Leamington Formation, the Goldson conglomerate and the Randels Cove conglomerate are treated together as interstratified units of demonstrable Ashgill age.

Definition

Several biostratigraphically distinct lenticles of Goldson conglomerate are interstratified with Helwig's (1967) Point Leamington Greywacke, and have been included within William's (1991) Point Leamington Formation. For purposes of description in this report, the term Randels Cove conglomerate is restricted to that polymict conglomeratic turbidite unit that regionally overlies the Point Leamington Formation (Dec, 1993). The Goldson conglomerate lenticles are, however, lithologically similar to the much thicker Randels Cove conglomerate.

Distribution and Thickness

Within the western part of the area surveyed, Late Ordovician siliciclastic turbidites outcrop in two separate structural basins; one centred around New Bay and the other located on the eastern side of the Fortune Harbour peninsula. The major lithological constituent of both basins is a thin- to medium-bedded sandstone turbidite succession belonging to the Point Leamington Formation. It contains a variably thick interval of siltstone turbidite and graptolitic mudstone dominated by mass flow sedimentary deposits.

As defined above, the Randels Cove conglomerate is entirely restricted to the type area of the Point Leamington Formation. It probably has a total combined thickness of at least 1 km in the central part of the New Bay basin. Thin lenticles of Goldson conglomerate are more common in the New Bay structural basin than on the eastern Fortune Harbour peninsula. Presumed Late Ordovician strata on nearby islands in the western Bay of Exploits are lithologically distinct, massive granular turbidites, which have been traditionally assigned to the Sansom greywacke (Oversby, 1967; Dean, 1977).

Lithology

In the South Arm–Osmonton Arm area, the Point Leamington Formation consists of two regionally mappable subunits (O'Brien, 1990, 1993). These are present in changeable proportions and varying thicknesses within different parts of the New Bay structural basin, although they occur everywhere within the same ascending sequence. The most widespread and volumetrically significant subunit of the Point Leamington Formation consists of interbedded grey sandstone and siltstone with some quartz-rich wacke and light-grey shale (Plate 23). This lithofacies is observed to be in primary stratigraphic contact with several Goldson conglomerate lenticles and with the base of the Randels Cove conglomerate (Williams *et al*., 1992). The sandstone turbidite subunit is most prominent in the lower part of the Point Leamington Formation.

The second lithological subunit of the Point Leamington Formation is composed of light-grey laminated siltstone and thin-bedded sandstone turbidite, dark-grey carbonaceous shale (with some debris flows) and olistostromal melange with a black shale matrix (Plate 10). The subunit is lithologically distinctive and contains a relatively rich and biostratigraphically diagnostic graptolite fauna (Williams, 1991; Williams *et al*., 1992). The subordinate, dark-coloured, mudstone-rich turbidites, which are positioned stratigraphically between the lowest and highest lenticle of Goldson conglomerate, form regional-scale splits in the lower sandstone turbidite division. A possibly equivalent, albeit much thinner, mudstone-rich turbidite subunit may be present in the Baptist Cove area on the Fortune Harbour peninsula.

Although there are several stratigraphically distinct bodies of Goldson conglomerate in the New Bay area, each wedge-shape lenticle contains interstratified and graded units of grey pebbly wacke and grey cobble- to boulder-conglomerate (O'Brien, 1990). Within most lenticles, pebbly wacke deposition preceded polymictic conglomerate deposition. Conspicuous detrital lithologies in polymictic conglomerate horizons include turquoise chert and pillow

Plate 23. *Re-amalgamated sand horizon (under the camera lens cap) in the type area of the Point Leamington Formation is overlain by a laterally-continuous graded bed with a locally cuspate base but without other modification of the original Bouma turbidite divisions. It is underlain by a sequence of graded turbidites which are displaced by domino-type listric normal faults that extend upwards to the quicksand.*

lava probably derived from the upper Exploits and upper Wild Bight groups, foliated tonalite very similar to that found in the South Lake Igneous Complex, and *N. gracilis* Zone black shale and grey mottled chert typical of the lower Lawrence Harbour Formation.

The Randels Cove conglomerate occupies a regional synclinorium within the New Bay structural basin and is observed to conformably overlie the Point Leamington Formation in several localities. Large rafts and conspicuous clasts of shallow-water coralline limestone have yielded Late Ordovician fossils that provide older age limits for the deposition of this thick conglomerate horizon (*see below*). The map unit is much older than the thick polymict conglomerate of the Goldson Formation of New World Island (Dec *et al*., 1993; Table 1).

The stratigraphic top of the Randels Cove conglomerate is not exposed in the New Bay area. The map unit is probably regionally discontinuous and restricted to the New Bay area however, as fine-grained Badger Group turbidite deposits of upper Ashgill and Llandovery age in the western Bay of Exploits and near the Exploits River estuary are not underlain by biostratigraphic equivalents of the Randels Cove conglomerate.

Biostratigraphy

The definition and graptolite biostratigraphy of the Upper Ordovician Point Leamington Formation was recently revised by Williams (1991). Where the upper Lawrence Harbour Formation underlies the first thick siliciclastic sandstone bed of the Point Leamington Formation, the transition between these lithostratigraphical units occurs at a level within the *Dicranograptus clingani* or *Pleurograptus linearis* zones, depending upon locality. Stratigraphically higher intervals of interbedded shale and sandstone within the formation have been demonstrated to belong to distinctly different graptolite biozones, and assemblages indicative of the *Dicellograptus complanatus* and *Dicellograptus anceps* zones of the Ashgill are present (Table 1).

More recently, ten new graptolite localities were discovered during systematic remapping of the Point Leamington Formation in the New Bay structural basin (Williams *et al*., 1992). The graptolite species present were similar to those described by Williams (1991) and the additional faunal assemblages simply confirm biostratigraphic conclusions about the regional representation of the *D. complanatus* Zone and *D. anceps* Zone. However, such new graptolite localities are useful in identifying and mapping certain stratigraphic intervals in the type area of the Point Leamington Formation.

Faunal assemblages indicative of the *D. complanatus* Zone occur near the base of the largest mappable lenticle of fossiliferous turbidites in the Point Leamington Formation. Extending from Strong Island Sound to the vicinity of Southwest Arm, this thinly stratified lenticle of interbedded dark grey shale and light grey concretionary sandstone is host to most of the olistostromal deposits in the Point Leamington Formation. Debris flows underlying Goldson conglomerate lenticles have disrupted blocks of concretionary siltstone turbidites that have yielded *D. complanatus* Zone graptolites. Although much smaller than the largest body, lenticles of dark grey mudstone-rich turbidite are located in Osmonton Arm near West Hare Island and Besom Cove. They are also probably present near Island Pond on the peninsula separating Osmonton Arm and Southwest Arm, although the lack of identifiable graptolites precludes their contemporaneity.

Strata of the *D. complanatus* Zone interval provide a younger age limit for several lithological subunits in the lowest 100 m to 1000 m of the Point Leamington Formation. Immediately underlying *D. complanatus* Zone strata is a regionally developed sequence marked by distinctively branched, pyritic, bedding-parallel, inorganic pseudofossils. The pseudofossil horizon is typically found in the lowest grey shale-dominant interval of the Point Leamington Formation, commonly hundreds of metres above the top of the Lawrence Harbour Formation.

The monotonous turbidite sandstone succession below this pseudofossil horizon contains several of the stratigraphically lower lenticles of Goldson conglomerate in the Point Leamington area. The lowest Goldson conglomerate lenticle occurs near the Point Leamington–Lawrence Harbour transition and, in places, this conglomerate is underlain by the formation's oldest known olistostrome. This particular debris flow was deposited directly above dark shale yet it is marked by distinctive limestone blocks. Lithologically similar to the detrital limestone clasts observed in the upper part of the adjacent Wild Bight Group, these carbonate olistoliths contain poorly preserved conodonts which may range from the Arenig to the Llandeilo (F.H.C. O'Brien, personal communication, 1992).

The generalized stratigraphy of the pre-*D. complanatus* portion of the Ashgill succession in the New Bay area is, in ascending order, thin- to medium-bedded turbidite sandstone, massive pebbly wacke and polymictic conglomerate, and siltstone–sandstone rhythmites with conspicuous pseudofossils. This sequence is capped by the previously discussed olistostrome-bearing shale-rich lenticles and floored by either the oldest (*D. clingani* Zone) grey shale lenticles in the Point Leamington Formation or the black shales of the Lawrence Harbour Formation. Characteristic of the marginal regions of the entire New Bay structural basin, and probably in part of *P. linearis* Zone age (cf. Williams 1991), the lowest part of the Point Leamington Formation varies considerably in total thickness, although a predictable sequential order of individual subunits is commonly present.

Several horizons within the fossiliferous shale-rich subunit of the Point Leamington Formation and below the uppermost lenticle of Goldson conglomerate contain graptolites belonging to a higher part of the *D. anceps* Zone (Rawtheyan stage of the Ashgill). From a biostratigraphic perspective, the most important specimens are *D. anceps* (Nicholson) and *Paraorthograptus pacificus* (Ruedemann). The latter species is characteristic of the *P. pacificus* Subzone of the *D. anceps* Zone in southern Scotland. In places, grey shale and siltstone interbeds within the Point Leamington Formation separate small tongues of Goldson conglomerate and illustrate Bouma intervals indicative of turbidity current sedimentation. They contain graptolites from the lower part of the *D. anceps* Zone (the *D. complexus* Subzone or the Cautleyan stage of the Ashgill).

The lenticle of mudstone-rich turbidites in the Point Leamington Formation that contains *D. complanatus* Zone assemblages near its base extends stratigraphically upward through the *D. complexus* Subzone and into the *P. pacificus* Subzone of the *D. anceps* Zone. Where it obtains its maximum 1 km thickness, this lenticle is host to five stratigraphically discrete olistostrome horizons (O'Brien 1990). Each of these debris flows is separated by unbroken Point Leamington Formation carrying in situ *D. anceps* Zone assemblages. The youngest dated block of transported material in these olistostromes is Ashgill in age, although Caradoc olistoliths are also present (Williams 1991). Slump folded shale blocks have yielded graptolites belonging to the *P. linearis* Zone, while rip-up clasts of nodular limestone contain conodonts indicative of the *Amorphognathus superbus–Amorphognathus ordovicicus* biozones (Williams *et al*., 1992). Both are consistent with the age of the adjacent unbroken strata in this part of the Point Leamington Formation.

In the vicinity of Western Arm, strata of the Point Leamington Formation overlying the large grey shale- dominant lenticle contain graptolite assemblages which also place these beds within the *D. anceps* Zone. They directly underlie the main body of the Randels Cove conglomerate. In the core of the regional syncline, the stratigraphically highest part of the Randells Cove conglomerate contains carbonate bioclasts which have yielded mid–late Ashgill conodonts (F.H.C. O'Brien, unpublished data) and a latest Ashgill shelly fauna (W.D. Boyce, unpublished data). Whereas the Goldson conglomerate lenticles formed in the Pusgillian and Cautleyan stages (early to mid Ashgill), the Randells Cove conglomerate may have been deposited in an interval ranging from the Rawtheyan stage to the top of the Hirnantian (in the late Ashgill).

Interpretation

The biostratigraphic information discerned from the graptolites, conodonts and shelly fauna is of fundamental importance in interpreting the lithostratigraphically isolated lenticles of pebbly wacke and conglomeratic turbidite within this Ashgill sedimentary basin. It successfully demonstrates that these are not contemporaneous tongues emanating from a single coeval sheet deposit of polymict conglomerate within a laterally-split sandstone turbidite package. Instead, lenticles belonging to several different stages of the Late Ordovician generally become thicker and more laterally continuous upward in the succession. Conglomeratic infilling of the basin, which occurred throughout much of Ashgill time, overlapped most of the period of accumulation of the host sandstone and siltstone succession.

Upper Black Island Greywacke

The term Upper Black Island greywacke is used informally in this report to describe one of the most areally restricted, lithostratigraphically unique and biostratigraphically complete units of the Badger Group.

Distribution and Thickness

The Upper Black Island greywacke occurs in several discontinuous tracts on the northeasternmost part of Upper

Black Island, the southern part of Hummock Island and on intervening islets in the western Bay of Exploits (Figure 3). It is also exposd near the community of Little Burnt Bay and along the coast west of Southern Head (O'Brien, 1992b). The largest area of contiguous exposures of this map unit measures approximately 0.4 km^2 on the islands, although undersea outcrop may be more extensive. The total exposed stratigraphic succession of the Upper Black Island greywacke is estimated to be less than 200 m in thickness.

Small fault-bounded tracts of the Upper Black Island greywacke are observed to be juxtaposed with the pebbly mudstone and broken formations of the Dunnage Melange, various units from the middle and upper Exploits Group, and the black shale and chert divisions of the Luscombe Formation. They also occur as hornfelsed septae within sheeted intrusions of gabbro and diorite near the margins of the Loon Bay batholith.

Stratigraphic Nomenclature

The various sedimentary strata in the Upper Black Island greywacke were previously assigned to Williams' Goldson Formation of the Botwood Group, although Williams (1962) also included limestones and basalts from other units on Upper Black Island in his Goldson Formation. In contrast, Oversby (1967) considered what is here termed the Upper Black Island greywacke to be a partial lateral facies variant of the Point Leamington greywacke. However, he also incorrectly correlated an argillite and shale-rich part of this Badger Group map unit with Helwig's (1967) Lawrence Harbour Shale of the Exploits Group (i.e., the northern fault block of Oversby's Caradoc black shale unit).

The Upper Black Island greywacke is also equated, in part, with the Upper Black Island argillite as mapped by O'Brien (1990, 1991b). Basaltic breccias previously included with O'Brien's (1991b) map unit of fossiliferous argillites have been reassigned to the Lawrence Head Formation of the Exploits Group, which is seen to be fault imbricated with the Upper Black Island greywacke.

Lithostratigraphy

The stratigraphic base of the Upper Black Island greywacke is drawn at the lowest dark-grey graded sandstone of an approximately 20 m thick, medium-bedded, sandstone turbidite sequence. The basal contact is gradational over 2 m. The underlying iron oxide-weathered, pyritiferous black siltstone and interbedded black carbonaceous shale may belong to the thin-bedded upper part of the Luscombe Formation. The stratigraphic top of the Upper Black Island greywacke is not observed in central Notre Dame Bay.

In ascending order, the Upper Black Island greywacke is composed of sandstone turbidites, carbonate olistostromes, chert breccias, graded wackes, marls or limestones, and siltstone turbidites.

Lithology

The quartz-rich sandstones at the base of the Upper Black Island greywacke are medium-bedded, graded and display incomplete sequences of Bouma intervals. The overlying carbonate sequence is massive or crudely stratified and mostly composed of a series of olistostromal deposits, each having variably sized limestone blocks set in a calcareous shale matrix. An interval of laminated red chert directly overlies the highest carbonate olistostrome. It is succeeded by thinly bedded grey cherts, siliceous grey argillites and minor red cherts interstratified with graded laterally continuous wackes. The wackes have erosive bases and pebble- to sand-size clasts.

Higher in the succession, poorly stratified chert breccias and thick-bedded, granular-to-pebbly, graded wackes dominate the Upper Black Island greywacke. However, the youngest preserved sequence begins with limy shales, calcareous sandstones, marls and thin discontinuous limestones, and passes upward into medium- to thin-bedded siltstone turbidites. These fine-grained siliciclastic beds contain conspicuous, dark grey, argillaceous partings.

Deformed gabbro sills are seen to intrude and thermally metamorphose some of the fossil-bearing strata within the Upper Black Island greywacke.

Biostratigraphy

In several locations, the basal turbidite sequence of the Upper Black Island greywacke is seen to conformably overlie a fossiliferous black shale sequence which is known to belong to the *D. clingani* graptolite biozone. Thin intervals of dark grey siltstone and shale in the sandstone turbidite beds above these upper Caradoc shales have yielded lower Ashgill (Pusgillian) graptolites and most probably belong to the *P. linearis* and *D. complanatus* zones (S. H. Williams, 1992, unpublished data).

In the one locality where the basal unit of the Upper Black Island greywacke is represented by a carbonate olistostrome, a spectacular block-in-matrix melange is observed to directly overlie carbonaceous black shales (Oversby, 1967). Graptolites recovered from the top of the black shale sequence immediately beneath the erosional disconformity belong to the *D. clingani* Zone and are correlatable with the black shale sequence seen beneath the Pusgillian turbidite succession. Some carbonate olistostromes in the Upper Black Island greywacke above the basal disconformity are possibly the along-strike equivalents of some of the *P. linearis* Zone and *D. complanatus* Zone sandstones, siltstones and shales.

Olistostromes in the Upper Black Island greywacke above the Pusgillian turbidites contain large olistoliths of shallow-water reefal limestone (Plate 24). During recent biostratigraphic sampling of this olistostromal sequence (W.D. Boyce, 1993, unpublished data), blocks noted as having a shelly fauna were also found to contain conodonts. The presence of *Amorphognathus* sp. and *Protopanderodus liripipus* indicate that at least some of the resedimented carbonate blocks are late Caradoc or Ashgill in age, and not Silurian (F.H.C. O'Brien, unpublished data). Light-coloured conodonts were discovered in several different olistoliths and the conodont alteration index (CAI) was determined to be as high as 7 in several of the marble blocks.

In earlier studies, some of these carbonate blocks had been reported to contain a Silurian shelly fauna and, on this basis, the debris flow-bearing unit was correlated with the Silurian Goldson Formation of New World Island (Williams, 1962). Shrock and Twenhofel (1939) were the first authors to record Llandovery favositid corals from the olistostromal unit in what is here termed the Upper Black Island greywacke. Boucot and Smith (1978) described a thermally metamorphosed marble from Upper Black Island which the author considers as an olistolith in one of the above mentioned debris flows. The marble is stated as having unique pentamerinid and atrypacean brachiopods that indicate a probable late Wenlock–Ludlow age but an unequivocal Late Silurian age assignment.

Plate 24. *An olistostrome in the Upper Black Island greywacke is predominantly composed of shelly limestone blocks (light colour) set in a limy mudstone matrix (dark colour). Locally scoured into* D. clingani *Zone black shales, only Ashgill conodonts have been recovered from the carbonate megablocks in the island's olistostromal melange deposits.*

Plate 25. *In certain parts of the Upper Black Island greywacke, interbedded limestone and siliciclastic wacke (below the hammer), together with younger cherts, form coherently bedded successions, which are situated stratigraphically above the limestone olistostromes on the northeast coast of the island.*

Marls and thin limestones occur higher in the succession of the Upper Black Island greywacke (Plate 25), although certain beds may possibly be the unbroken equivalent of some of the carbonate olistostromes. Some limestone beds have yielded a diverse shelly fauna (Boyce *et al*., 1991), including paleogeographically and paleoenvironmentally significant cyclopygid trilobites of late Ashgill age (Plate 26).

Abundant mid-Llandovery graptolites have been recovered from the uppermost exposed argillaceous turbidites of the Upper Black Island greywacke. The graptolite taxa present are diagnostic Silurian forms, including *Rastrites peregrinus* (Barrande), *Monograptus austerus sequens* Hutt, *Orthograptus insectiformis* (Nicholson), *Monograptus* spp. indet, *Climacograptus*? sp., and *Glyptograptus*? sp. (Williams and O'Brien, 1991). These graptolite-bearing argillites must fall somewhere between the *M. triangulatus* and *M. convolutus* zones of the Lower Silurian sequence in Britain. It may thus be stated with certainty that the highest turbidite sequence in the Upper Black Island greywacke is Aeronian in age (Table 1).

Assuming that the writer has correctly restored the stratigraphic section of the Upper Black Island greywacke and that the above graptolites are correctly identified, the

Plate 26. *Slump-folded and pulled-apart limestone beds occur within the sandy and argillaceous turbidites of the late Ashgill–Llandovery succession on northeastern Upper Black Island. The illustrated carbonates have yielded Rawtheyan cyclopygid trilobites.*

postulated age of the pentamerind and atrypacean brachiopods recovered from the marble olistostrome is at least 10 My too young (Boucot and Smith, 1978; Table 1).

Interpretation

The Late Ordovician and Early Silurian ages of the various hemipelagic, siliciclastic and limy turbidite deposits comprising the Upper Black Island greywacke are coeval with and considerably younger than the strata in the adjacent Point Leamington Formation. The Upper Black Island greywacke is, however, generally lithologically distinct and much thinner than the Point Leamington Formation.

The Upper Black Island greywacke is the longest ranging and thinnest of the known units in the Badger Group of central Newfoundland. If the early Ludlow age of certain limestone olistoliths in melange is substantiated by further work, the temporal development of the Upper Black Island greywacke may partly overlap that of the shallow-marine Indian Islands Group, located to the east of the Dog Bay Line (Figure 3).

Campbellton Greywacke and Lewisporte Conglomerate

The terms Campbellton greywacke and Lewisporte conglomerate are used informally in this report to describe two regionally mappable subunits of the Badger Group in the southeastern part of the study area. Together, they comprise a distinctive sequence of lithofacies associations that have repeatedly recurred in both Ashgill and Llandovery times.

Distribution and Thickness

Units of Campbellton greywacke and Lewisporte conglomerate crop out in two discrete areas separated by the younger terrestrial strata of the Botwood Group (Figure 3). In the north, they occur within a fault-imbricated assemblage along the southern coast of the Bay of Exploits, where individual Badger Group tracts typically vary between 30 to 60 km^2 in outcrop area. Farther south, in the vicinity of Notre Dame Junction, another regional fault-bounded tract of the Badger Group is located near the margin of the Mount Peyton batholith (Figure 3).

Estimations of unit thicknesses are difficult to calculate due to the abundance of periclinal folds within individual fault-bounded tracts. However, the combined structural and stratigraphic thickness of the smallest fault-bounded tract of the Campbellton greywacke and Lewisporte conglomerate is approximately 400 m. The largest fault-bounded tract is estimated to be approximately 2400 m thick.

The Campbellton greywacke and Lewisporte conglomerate units display similar geometries as the Goldson conglomerate lenticles in the Point Leamington greywacke. However, compared to the New Bay area, the greywacke and conglomerate units in the southeastern part of the Bay of Exploits are interstratified on a regional scale and the conglomerate tongues of the Badger Group are generally much larger in size.

Lithostratigraphy

The stratigraphic base of the Campbellton greywacke is exposed near Shoal Point in Lewisporte Harbour. Its stratigraphic top is, however, not encountered within the study area.

The Campbellton greywacke displays two fundamental lithodemic associations that are interstratified on various scales. An association of altered and partially replaced, fine-grained hemipelagic rocks is well developed around Lewisporte where, volumetrically, it represents most of the Campbellton succession. In the type area of Campbellton Harbour, the hemipelagic strata occur at the stratigraphic base of the

Campbellton greywacke succession and occur immediately south of the upper Caradoc black shales exposed in the Old Iron Mine (Figure 16).

An association of unaltered, coarser-grained siliciclastic rocks is the more widespread of the two lithodemic groups in the Campbellton greywacke, especially if the entire region between Indian Arm and Norris Arm is considered as a whole (Figure 3). In most localities in this area (e.g., Figure 16), sequences of sandy to granular wackes appear to be preferentially developed above variably thick hemipelagic successions, although the hemipelagites probably differ in age from one area to another.

Lenticles of Lewisporte conglomerate are interstratified within both the hemipelagic-dominated and siliciclastic-dominated successions of the Campbellton greywacke. These polymict conglomeratic turbidites have a highly variable thickness and lateral extent.

Lithology of the Campbellton Greywacke

The Campbellton hemipelagic association is typified by variably silicified, iron enriched, thin-bedded and laminated units of grey-green argillite interstratified with grey-green shale. More conspicuous are light-green shale and dark-green argillite rhythmites, mottled and slumped-folded laminites made up of dark-grey, black and green mudstones, and interlaminated light-grey and light-green cherty argillites (Plate 27). Alternating dark-green and dark-red intervals of sandstone grading to siliceous argillite and locally associated beds of chert-pebble microconglomerate comprise the most readily recognizable elements of the Campbellton hemipelagic association. This association also includes some minor debris flows containing small blocks of slump-folded dark-grey siltstone and grey laminated chert along with distinctive clasts of recrystallized limestone and rounded granitic rocks.

The Campbellton siliciclastic association is characterized by massive to thick-bedded, graded and flute-marked, pebbly-, granular- and sand-sized wackes that are scoured into subordinate, light-grey, medium-bedded, nodular sandstone turbidites. In proximity to siliceous hemipelagic strata at the base of Campbellton greywacke, graded siliciclastic sandstones displaying Bouma intervals develop a distinctive pink hue as a result of secondary mineralization.

Regionally mappable lenticles of thin-bedded dark grey siltstone and other shale-rich turbidites are notably absent in the Campbellton greywacke, which stands in marked contrast to the graptolite-bearing strata in the early–mid Ashgill part of the Point Leamington Formation. Fossil-bearing strata in the Norris Arm area include thin- to medium-bedded, calcareous sandstone turbidites, thin-bedded grey argillites, and graded pebble- to cobble- sized conglomerates (Plate 28).

Lithology of the Lewisporte Conglomerate

The lenticles of Lewisporte conglomerate are composed of grey, green and red polymict conglomerate with distinctive boulders and cobbles of jasper, metatonalite, turquoise chert and porphyritic basalt (Plate 29). Poorly stratified pebbly wackes containing similar detritus locally enclose outsized megarafts of coralline limestone up to 15 m in diameter (Plate 30).

Biostratigraphy of the Campbellton Greywacke

Fossiliferous strata in hemipelagic and siliciclastic facies of the Campbellton greywacke belong to several different stages of the Ashgill and the Llandovery. Near Shoal Point, at the stratigraphic base of the unit, a few specimens of the graptolite *Dicellograptus flexuosus* Lapworth suggest that the transitional interval between the basal Campbellton greywacke and the underlying black shale formation is latest Caradoc in age. The transitional beds occupy either the late *D. clingani* Zone or the *P. linearis* Zone (Williams 1993).

In the type area, where the Campbellton greywacke is underlain by the Luscombe Formation, slight exposure gaps are present near the equivalent horizon. The biostratigraphic data collected at Lewisporte's Shoal Point and Campbellton's Old Iron Mine probably indicates, though does not necessarily prove, that the thin-bedded interval of green siliceous argillites and siltstone turbidites overlying the transition zone at the base of the Campbellton greywacke belongs, in part, to the *P. linearis* Zone of the early Ashgill (Figure 16).

South of Lewisporte and west of Indian Arm Brook, several green bed and grey bed lithodemes typical of the hemipelagic association crop out extensively within a predominately southwest-facing succession of the Campbellton greywacke. In one locality, thin-bedded light-grey siliceous argillite is interstratified with discontinuous layers of chert pebble conglomerate and succeeded by mottled laminites of black and green mudstone. Boyce and Ash (1992) recovered a brachiopod–gastropod–trilobite–crinoid fauna from the pebbly chert interval. In the opinion of D.A.T. Harper (personal communication, 1993), the fauna is indicative of an uppermost Ashgill age (probably Hirnantian; Table 1).

A considerable thickness of unfossiliferous southwest-younging sandstone turbidites overlies these probable Hirnantian beds. So, it is quite possible that the Campbellton greywacke is as young as early Llandovery where it is overthrust by the terrestrial Botwood Group. It seems that, from its stratigraphic base near Shoal Point in Lewisporte Harbour southwards to its above-mentioned structural top, the Campbellton greywacke unit ranges through more of the Ashgill epoch than does the Point Leamington Formation.

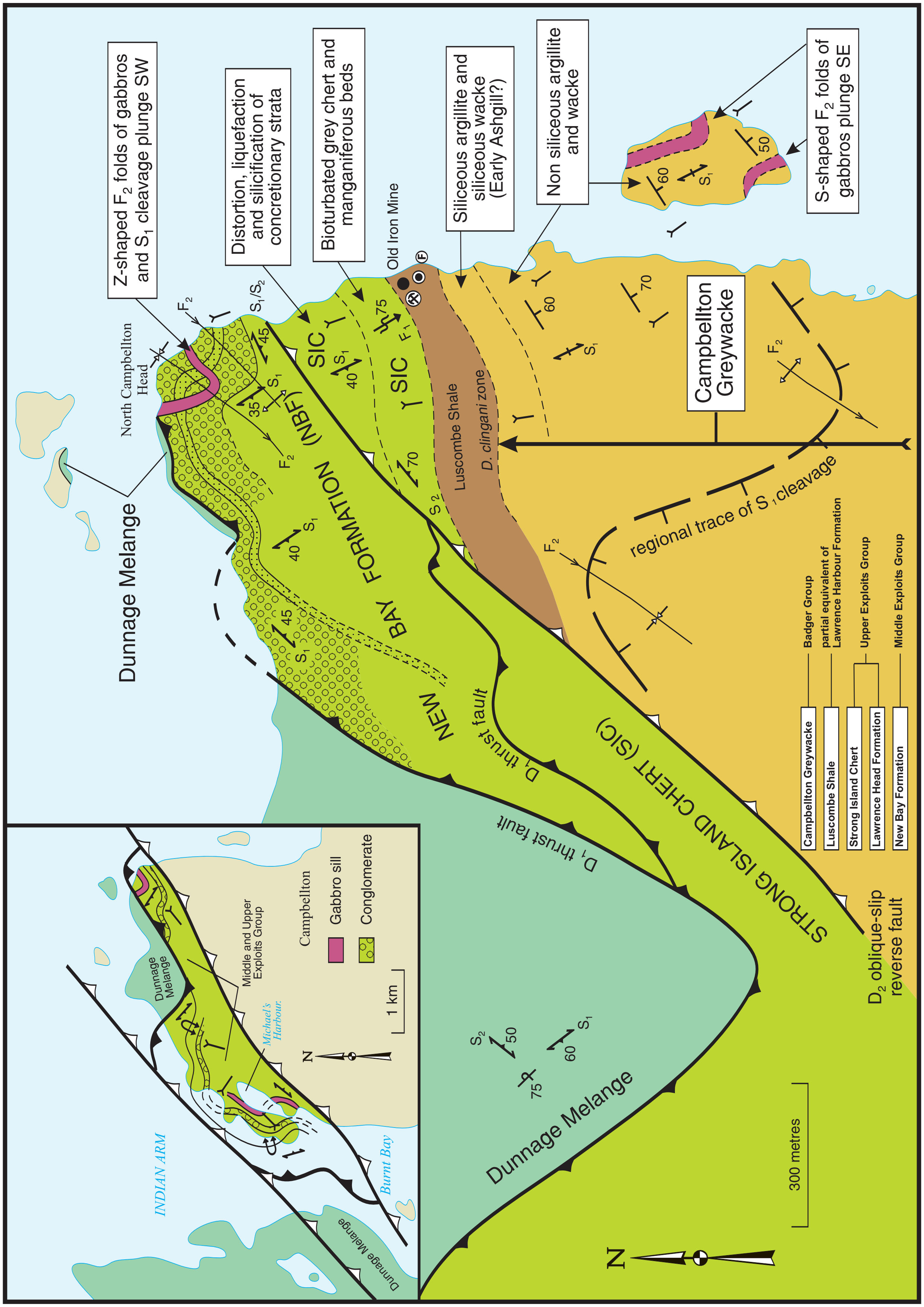


Figure 16. *Geological sketch map of the North Campbellton area illustrating some minor D_2 structures associated with a major southwest-plunging F_2 fold and a southeast-dipping D_2 reverse fault. Large- and small-scale D_1 structures are overprinted and are locally reoriented from a northwesterly strike into the northeasterly D_2 trend. Inset shows folded D_1 structures in an Exploits Group-defined D_2 structural basin that is bounded by D_2 reverse faults.*

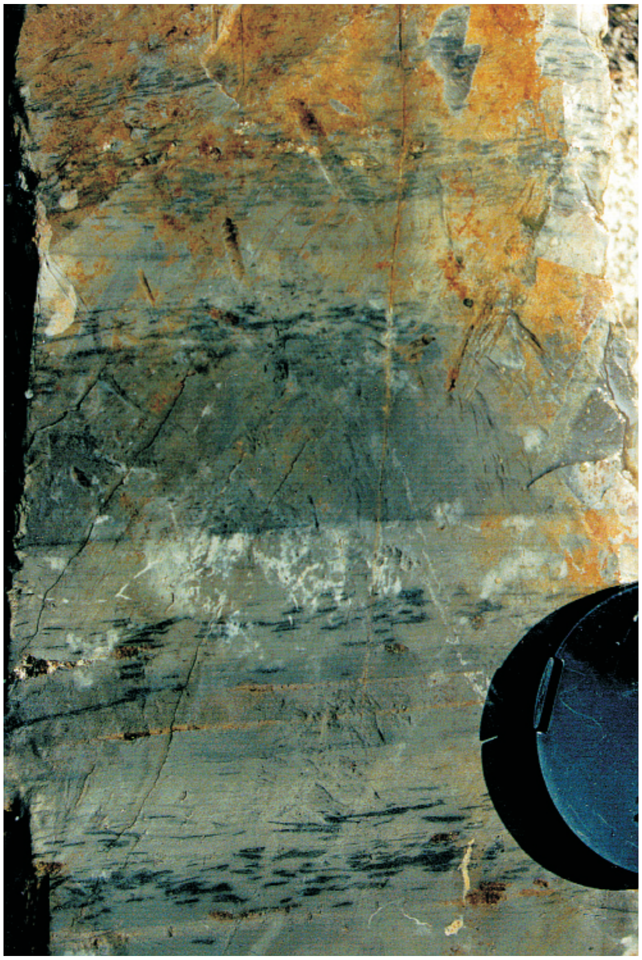


Plate 27. *Near Island Pond, a thinly bedded, sharply based, dark-grey siltstone grades upward into a laminated, light-grey, siliceous argillite marked by discontinuous laminae of black mudstone. Representing some of the associated hemipelagic lithodemes in the Campbellton greywacke, the original mud laminae in these turbidites (below lens cap) appear to have been physically disrupted prior to their silicification.*

Plate 28. *Thinly bedded, fine-grained, light-grey sandstone turbidites at Martin Eddy Point are interstratified with pebble conglomerates carrying rafts from subjacent sandstones. Representing some of the associated siliciclastic lithodemes of the Campbellton greywacke, such beds have yielded a late Llandovery shelly fauna on the Exploits River and farther southeast.*

The southern tract of the Campbellton greywacke and Lewisporte conglomerate is well exposed in the Norris Arm–Jumpers Brook region, where it is mainly composed of thin-bedded siliceous argillite, fine-grained and medium-grained sandstone turbidite and well-rounded pebble conglomerate. Although these Badger Group beds structurally overlie the terrestrial Botwood Group, they are lithologically similar to many of the turbidites described from the Badger Group panel structurally beneath the Botwood Group. They are also sporadically fossiliferous and locally contain a transported shallow-water shelly fauna (Boyce *et al*., 1994).

The fault-bounded turbidite tracts in the Norris Arm–Jumpers Brook region have been assigned to the Campbellton greywacke for several reasons. First, the undated oldest-exposed turbidites and siliceous argillites stratigraphically underlie a highly fossiliferous Upper Llandovery interval ranging from the latest Aeronian to the mid Telychian stage (Boyce *et al*., 1993; Table 1). Second, the lithological resemblance is more with the Campbellton greywacke than with strata in the lower–middle Point Leamington Formation, the lower Gull Island Formation, the upper Sansom Formation or the mid-upper Upper Black Island greywacke. Finally, several polymict conglomerate lenticles similar to the grey and red bodies of Lewisporte conglomerate are present in the Norris Arm area and are interstratified with the finer grained turbidites.

In summary, certain strata presently assigned to the Campbellton greywacke are younger than the mid Aeronian argillaceous turbidites of the Upper Black Island greywacke. They also considerably postdate all of the sandstone turbidites in the Point Leamington Formation. However, paleontological evidence of older Llandovery strata or corroboration of the presence of the Hirnantian and older Ashgill succession has yet to be found in the Norris Arm tract of the Campbellton greywacke.

Biostratigraphy of the Lewisporte Conglomerate

The lenticle of polymict conglomerate that immediately overlies the *P. linearis* Zone strata in the basal Campbellton greywacke sequence at Shoal Point crops out in the southern part of the town of Lewisporte. It may be the equivalent of the pre-*D. complanatus* Zone lenticle of Goldson conglomerate in the New Bay structural basin. Reports of Early Silurian corals from limestone clasts (Williams, 1962) in this lenticle of Lewisporte conglomerate would, however, point instead to correlations with units as young as or younger than the Randels Cove conglomerate.

Plate 29. *Lewisporte polymict conglomerate displays well-rounded, clast-supported cobbles and boulders of plutonic, volcanic and sedimentary rocks, as do other Goldson lithotypes. This particular conglomeratic lenticle of the Badger Group, which is thought to be younger than Hirnantian and older than late Telychian, is exposed at the High Point of Norris Arm near the fault contact with the Wigwam Formation of the Botwood Group.*

The coral-rich and echinoid-bearing limestone clasts in the various lenticles of Lewisporte conglomerate have not been systematically studied and the full biostratigraphic range of the conglomerate units is unknown. However, a long age range is probable, as the oldest conglomerate lenticle conformably overlies earliest Ashgill strata (Lewisporte Harbour) and the youngest conglomerate lenticle conformably underlies late Llandovery strata (Exploits River southwest of Wigwam Point).

Correlation

The northern tracts of the Campbellton greywacke are partly correlatable with some of the sedimentary strata that were earlier grouped in the Riding Island greywacke of the Campbellton sequence (Kay, 1975; Lash, 1994). The southern tracts of the Campbellton greywacke were previously assigned to the upper terrestrial formation of the Botwood Group (Williams, 1962) and they have never been formerly correlated with any unit of the Badger Group.

The grey chert–siliceous argillite sequence in the Upper Black Island greywacke is lithologically similar to certain intervals in the Campbellton greywacke. However, the thin-bedded hemipelagic strata in the Campbellton greywacke are orders of magnitude thicker and they do not contain the intercalated chert breccias or carbonate olistostromes typical of the Upper Black Island greywacke on Hummock Island.

Lithologically, the Lewisporte conglomerate resembles parts of the Goldson conglomerate, the Randels Cove conglomerate and the Goldson Formation. Some lenticles in the area of Lewisporte Harbour are probably equivalents of the early–mid Ashgill lenticles of Goldson conglomerate in the New Bay area. The region's thickest and most extensive lenticle of Lewisporte conglomerate occurs north of Island Pond, where it immediately underlies chert pebble conglomerate beds yielding a latest Ashgill shelly fauna. This large lenticle of Lewisporte conglomerate is thus possibly coeval with the Randels Cove conglomerate. Smaller size lenticles in the Norris Arm–Martin Eddy Point area may correlate with parts of the Goldson Formation on New World Island, although they are generally much thinner bodies of polymict conglomerate.

Interpretation

A Llandovery age for the coral-bearing conglomerate lenticle in Lewisporte Harbour would indicate localized condensation of the entire Ashgill portion of the Campbellton greywacke. Moreover, it would point to a stratigraphic record analogous to that documented in the Upper Black Island greywacke of the Badger Group. Alternatively, an early–mid Ashgill age for the limestone-bearing conglomerate would strengthen correlations with the Point Leamington Formation and indicate a much faster rate of sediment accumulation as seen in the New Bay area.

Lenticles of Lewisporte conglomerate that occur stratigraphically higher in the Campbellton greywacke occur farther south toward the Botwood Group. In places, this fossil-bearing part of the Ashgill section is faulted against the Llanvirn–Llandeilo Strong Island chert, the Caradoc Lus-

Plate 30. *Fossiliferous limestone clast, typical of those from the various lenticles of the Lewisporte conglomerate. The specimen illustrated is from a giant limestone block, measuring about 5 m x 15 m in dimension, set in conglomerate on an island in Southwest Pond.*

combe Formation and the lowest known polymict conglomerate lenticle. Thus, the regional interlayering of Lewisporte conglomerate and Campbellton greywacke tracts is partly the result of fault imbrication and structural duplication of the stratigraphy preserved in some of these Badger Group sequences. However, in other localities, it is also partly due to the primary interstratification of dated Late Ordovician and Early Silurian rock units.

In this regard, the biostratigraphic position of the fault-bounded lenticle of Lewisporte conglomerate outcropping between White Rock Pond and Indian Arm Brook is in doubt, as is the stratigraphically underlying tract of Campbellton greywacke.

Regional Interpretations

In Notre Dame Bay, the tectonic framework of Late Ordovician and Early Silurian deep-marine sedimentation is controversial, although most workers agree that basin fill generally coarsened-upward and that deposition accompanied active tectonism. Arnott *et al*. (1985) suggested that the eastern margin of the Notre Dame Subzone (by then an accreted part of composite Laurentia) became the site of a continent–ocean transform fault. Second-order wrench faults were envisaged to control Badger Group deposition in localized basins in the westernmost part of the Exploits Subzone overstep sequence. As an alternative model, van der Pluijm (1986) interpreted all of the Late Ordovician and some of the early Silurian turbidite tracts in the overstep sequence to represent fossiliferous offscraped successions in an accretionary prism dominated by southeast-directed thrusts. In this regional interpretation, most of the underplated rocks were adhered to the base of the Taconic Notre Dame arc (van der Pluijm and van Staal, 1988; van Staal *et al*., 1990), to the east of which lay the remnant seaways of the contracted Iapetus Ocean (Pickering *et al*., 1988).

Around most of the margin of the New Bay structural basin, laterally discontinuous lenticles of polymictic conglomerate and pebbly wacke are characteristic of a variably thick succession that records the interval between the *D. clingani* and *D. complanatus* graptolite biozones. In several locations, the Early Ashgill wackes that lie tectonically adjacent to South Lake tonalite contain abundant euhedral clasts of blue quartz and rare prisms of fresh hornblende in addition to pebbles of tonalite, granodiorite, turquoise chert and felsic volcanic rocks. This implies insignificant chemical weathering of the clasts' source area and little mechanical breakdown of the wackes' detrital constituents. As a nearby provenance is likely to explain the preferential erosion of the ubiquitous tonalite clasts in the early Ashgill conglomerate, the adjacent body of South Lake blue-quartz tonalite would be a reasonable source of such locally transported detritus. Furthermore, these observations suggest that the South Lake Igneous Complex and the supracrustal rocks of the Wild Bight Group were already in close proximity to each other by latest Caradoc time, with both supplying detritus to the Point Leamington sedimentary basin.

It is evident that at least one of the accumulation sites of polymictic conglomerate was initially controlled by debris flow-related slumping and mass wasting within a specific submarine channel, peculiar to one of the margins of the Point Leamington basin. Ashgillian uplift and resedimentation of Caradoc black shale from the Lawrence Harbour Formation occurred during the *D. complanatus* and *D. anceps* intervals (Williams *et al*., 1992). As discussed above, uplift of that portion of the basin margin which abuts the South Lake Igneous Complex may have occurred even earlier during the *P. linearis* Zone.

Faulting during the Ashgill–Llandovery

The New Bay Fault, which separates the latest Ashgill turbidites of the Randels Cove conglomerate from the earliest Arenig island-arc basalts of the Tea Arm Formation, is a representative regional fault that may have undergone growth during the mid-Paleozoic. Within this fault zone, Ashgill strata of the *D. complanatus* Zone are separated from uppermost Arenig–lowermost Llanvirn strata of the *D. hirundo* or *D. artus* biozones by a multi-unit sequence of folded and faulted beds whose combined structural/stratigraphic thickness is as little as 500 m. Of note is the fact that, some distance from the New Bay fault zone, the pre-*D. complanatus* portion of the Point Leamington Formation alone exceeds this thickness. Nevertheless, within this tract of attenuated stratigraphy, most regionally developed Ordovician units are present, such as the *Holmograptus*-bearing beds of chert and alkali basalt within the upper part of the Exploits Group, the *N. gracilis, C. bicornis*, *D. clingani and P. linearis* intervals from the lower, middle and upper Lawrence Harbour Formation, the lowest polymictic conglomerate lenticle of the Goldson conglomerate within the *P. linearis* Zone, and the distinctive pre- *D. complanatus* Zone pseudofossil horizon in the lower Point Leamington Formation.

The Middle and Late Ordovician succession within the New Bay fault zone is relatively thin and is interpreted to be stratigraphically condensed. In contrast, strata within the lower Point Leamington Formation west of the fault zone represent a much thicker succession but record a much shorter age span. This is possibly a consequence of early Late Ordovician growth faulting near the margin of the New Bay basin associated with renewed structural movements on the edge of the northeastern depocentre of the Exploits Group (*see* section on Faulting During Accumulation of the Exploits Group, page 37).

Syndepositional tectonism may have not only caused thickening of several Ashgill biostratigraphic zones from basin margin to basin centre, but it may also account for the exhumation of the youngest part of the South Lake Igneous Complex, the Pennys Brook Formation of the Wild Bight Group and the lower part of the Lawrence Harbour Formation. The regional juxtaposition of early Ashgill and late Arenig map units can not be solely explained by anomalously large, post-Ordovician displacements along the New Bay fault zone.

Other Controls on the Badger Group Basin

Any explanation for the development of the Badger Group in central Notre Dame Bay must explain the condensation of the entire group on Upper Black Island and the generation of thick siliciclastic lithofacies at the expense of carbonate and chert lithofacies in the Badger Group elsewhere. The Point Leamington Formation, the Upper Black Island greywacke and the Sansom greywacke each lie south of and tectonically adjacent to the Lukes Arm Fault; however, the lithofacies variation amongst such units occurs along strike from northwest to southeast and from southwest to northeast. Thus, in the study area, facies belts in the Badger Group are not simply controlled by distance from a Late Ordovician–uplifted block north of the Red Indian Line.

A critical feature of the map area is that Late Ordovician and Early Silurian deep-sea basins are not regionally coincident with the depocentres of the Early and Middle Ordovician island-arc basins. Some eustatically controlled transgressions and regressions recorded in Late Ordovician and Early Silurian strata of the Appalachian–Caledonian orogen are the right age (Fortey, 1984) to have affected the basin fill of the Badger Group overstep sequence in central Notre Dame Bay. Thus, a drop in sea level associated with the Late Ordovician glaciation of Gondwana is an alternative explanation for the destruction of Ashgill reefal carbonates, the erosion of early Ashgill and older marine strata, and the redeposition of such beds at different times in various basins of the Badger Group (Dec *et al*., 1993). However, the stratigraphic effects related to these global oceanic events are orders of magnitude larger in scale than those affecting the lithofacies of an individual biozone in any Badger Group lenticle or formation.

Although the plate tectonic cause of the multi-directional shift of these uniquely architectured depocentres is unknown, early–mid Paleozoic subsidence and deposition were episodic and unsteady rather than being continuous and site specific. The onset of this type of tectonic pattern within the marine Badger Group may herald the development of the more areally restricted terrestrial tracts in the overstep sequence, as represented by the Silurian Botwood Group.

Botwood Group

The Botwood Group comprises the youngest unit of stratified rocks in the overstep sequence of the Exploits Subzone. The type area of the Botwood Group occurs in the southern part of the study area, where the group is well exposed in coastal sections.

Distribution and Thickness

The Botwood Group crops out over an approximate 650 km^2 area between the Northern Arm and Long Reach faults (Figure 3). An estimate of the stratigraphic thickness of the entire Botwood Group is in the order of 2.5 to 3.5 km thick. The minimum stratigraphic thickness of the lower Botwood Group is about 1 km; whereas, the maximum stratigraphic thickness of the upper Botwood Group is estimated at approximately 2.5 km thick.

Stratigraphic Nomenclature

In the type area, the Botwood Group is composed of two formations, a lower volcanic unit named the Lawrenceton Formation and an upper sedimentary unit called the Wigwam Formation. The stratigraphically gradational contact between these formations is observable near Kite Cove and Burnt Arm Pond. Neither the stratigraphic base nor the stratigraphic top of the Botwood Group is seen in the study area. However, farther east in Notre Dame Bay, the lower stratigraphic boundary of the Botwood Group is seen where the Lawrenceton Formation stratigraphically overlies the Ashgill–Llandovery Stoneville conglomerate of the Badger Group (Williams, 1993).

As originally defined, the red and grey, coarse-grained, well-rounded, siliciclastic conglomerate of the Goldson Formation was included as the basal formation of the Silurian Botwood Group (Williams, 1962; 1972). However, in 1977, Dean informally dropped the marine Goldson Formation from his terrestrial Botwood Group. A considerable thickness of sedimentary rocks previously grouped with the Goldson and Wigwam formations were reassigned to the Badger Group during the present geological mapping of the study area. As employed in this report, the term Botwood Group refers only to the terrestrial strata assigned to the Lawrenceton Formation and the Wigwam Formation.

Plate 31. *Flow-banded red rhyolite of the Lawrenceton Formation of the Botwood Group displays flow folds (left of hammer) in the Laurenceton–Burnt Arm Pond area.*

Small outliers of subaerial felsic volcanic rocks, such as those in the Charles Lake volcanic unit (Swinden, 1988; Dickson, 1998) and the Stony Lake volcanic unit (Colman-Sadd and Russell, 1982), lie above some of the presumed Early Ordovician rocks of the Exploits Subzone. They are known to be Early and Late Silurian (Dunning *et al*., 1990; Dunning, unpublished report, 1999), but have ambiguous relationships with the dominantly mafic volcanic rocks of the Lawrenceton Formation of the Botwood Group.

The Lawrenceton Formation is well exposed in the spectacular coastal section near the community of Laurenceton, although only the upper part of the unit is represented. The lower and upper subunits of the Wigwam Formation are present in the type section near Peterview; however, they are not observed in their original stratigraphic order. The fossiliferous Badger Group crops out near Wigwam Point, not the Botwood Group (Shrock and Twenhofel, 1939).

Lithostratigraphy

The Lawrenceton Formation of the Botwood Group is a bimodal mafic–felsic volcanic unit, dominated by highly vesicular basalt flows. Scoriaceous basalt and banded ignimbrite are representative subaerial volcanic facies of the Lawrenceton Formation (Plates 31 and 32). The Wigwam Formation of the Botwood Group consists of grey, green and red, parallel-laminated and crossbedded sedimentary strata. Red micaceous sandstone, which forms the youngest and thickest part of the formation, is a characteristic subaerial sedimentary facies (Plates 33 and 34).

Internal subunits of the Lawrenceton and Wigwam formations are widespread throughout the Exploits River–Bay of Exploits region (O'Brien, 1993a; Dickson and Colman-Sadd, 1993; Dickson *et al*., 1995). Each formation is informally separated into subunits that are mappable on 1:50 000 scale geological maps of the central Notre Dame Bay area (e.g., Dickson *et al*., 1995).

Lithology and Stratigraphy of the Lawrenceton Formation

The lowest exposed part of the Lawrenceton Formation is made up of poorly stratified, grey, crystal-lithic tuff and coarse felsic breccia interstratified with minor basalt flows. In places, highly sheared felsic and mafic rocks are sericitized and pyritized (Dickson and Colman-Sadd, 1993; Colman-Sadd, 1994). The lower part of the formation, which measures less than 0.5 km in total thickness, also contains grey aphanitic intermediate tuffs, minor lahar units displaying large blocks of ignimbrite and basalt, and pink or grey flow-banded rhyolite (Plate 31).

The upper part of the Lawrenceton Formation, at less than one kilometre thick, has more diagnostic rocks than the lower part. These include extremely vesicular, purplish-grey porphyritic basalts and reddish-purple autobrecciated basalts (Plate 32). Some of the stratigraphically highest basalts in the Lawrenceton Formation have deep crevasses developed near their vesiculated flow tops and are filled with red sandstones displaying primary geopetal features (Plate 35). Discontinuous bands of interflow red conglomerate are especially well developed near the top of the Lawrenceton Formation, where they display abundant clasts of locally derived basalt.

Plate 32. *Vesicular basalt from the upper part of the Lawrenceton Formation near Kite Cove displays a reddened flow-top breccia cemented by grey–green sandstone. Stratigraphic top of the basalt is toward the end of the hammer handle.*

Lithology and Stratigraphy of the Wigwam Formation

The overlying Wigwam Formation of the Botwood Group has grey, medium-bedded, trough cross-stratified sandstone at its base. In places, thin- to medium-bedded intervals of conspicuous quartz-rich sandstone are present within grey bed sequences that are up to 1 km thick. Particular horizons are also notably rich in detrital mica. Some of the ripple-marked tops of the grey sandstones illustrate polygonal mudcracks and presumably indicate desiccation within an arid estuarine or fluvial environment. The lowest Wigwam Formation sandstones are locally interstratified with thin discontinuous polymict conglomerates. Above a Lawrenceton Formation inlier south of Peters River Pond, such sandstones display detrital clasts of Campbellton greywacke.

Higher in the lower Wigwam succession, green and grey trough cross-stratified sandstones are interbedded. At the top of this subunit, red and green sandstones are interstratified over several hundred metres and contain minor intervals of red slate and laminated argillite. Orange-weathered sandstones from the lower Wigwam Formation (Plate 36) were originally grey sandstones that have been reddened by the addition of pyrite and secondary ferroan carbonate.

A thick red bed sequence occurs in the upper part of the Wigwam Formation. Composed of some 1500 m of trough cross-stratified, ripple-marked, mud-cracked, massive to thick-bedded, red micaceous sandstone (Plate 33), it is the most widespread subunit of the Wigwam Formation. Intraformational microconglomerate in red sandstone shows an abundance of red shale fragments. Buff varieties of sandstone in this Wigwam subunit represent original red beds that were altered and leached near faults.

Thin intervals of grey sandstone are locally developed near the base of the upper red sandstone subunit within the town of Botwood. Rare basaltic lava flows are present in the upper part of the formation south of the community of Laurenceton. Mafic volcanic rocks younger than the Wigwam Formation red beds have also been reported in the Botwood Group on the Change Islands (Currie, 1995a).

Age

The only fossils recovered from the Botwood Group are found in the lower part of the red bed succession of the upper Wigwam Formation northwest of Munroe's Pond. They are stratigraphically located some 400 m above the top of the Lawrenceton Formation, although the fossils do occur near an isolated basalt flow. Bivalves winnowed into interripple troughs indicate a probable late Ludlow age for this part of the Wigwam Formation (Boyce and Ash, 1994). On the Port Albert peninsula, Elliot *et al*. (1991) argued that at least part of the Wigwam Formation must be older than a crosscutting dyke dated at 422 ± 2 Ma (latest Wenlock–earliest Ludlow in Table 1).

Felsic volcanic flows within the Lawrenceton Formation have low zircon concentrations that distinguish them from dated latest Llandovery–Wenlock rhyolites in the Springdale Group and other Silurian terrestrial basins in Newfoundland. The Lawrenceton Formation could not be older than the uppermost Llandovery (upper Telychian), if it is assumed that the fauna from the Norris Arm tract of the Campbellton greywacke indicate a younger depositional age limit for the Badger Group and an older depositional age limit for the Botwood Group.

The terrestrial formations of the Botwood Group might have been deposited at any time between the latest Llandovery (Early Silurian) and the late Ludlow (Late Silurian).

Correlation

Rocks of the Botwood Group are best correlated, lithostratigraphically, with the terrestrial volcanic and sedimentary strata of the Springdale Group (Figure 3). The Silurian Springdale Group belongs to the overstep sequence of the Notre Dame Subzone.

An elongate body of red beds (the Rogerson Lake conglomerate of Colman-Sadd and Russell, 1982; Table 1) extends to the south and west of the main volcanosedimentary basin of the Botwood Group. These strata rest nonconformably on late Precambrian intrusive rocks which may represent the basement of the Exploits Subzone. A similar body of conglomerate occurs along the trace of the Chanceport Fault to the north and east of the Springdale Group (Coyle and Strong, 1987). There, red-bed conglomerate unconformably overlies Middle Ordovician and older rocks of the Roberts Arm Group of the Notre Dame Subzone (Figure 3).

The Botwood Group may possibly have some biostratigraphic equivalents within the lower formations of the Indian Islands Group (Table 1).

Plate 33. *Topset truncation of foresets in red, cross-bedded, micaceous sandstones at Winter House Cove. Typical of strata found in the upper part of the Wigwam Formation of the Botwood Group, the more recessive foresets are covered with darker coloured mud drapes.*

Regional Interpretations

In parts of central Newfoundland, strata of shallow marine and terrestrial origin are thought to have been deposited without significant hiatus in a shoaling-upward Silurian succession (Williams *et al*., 1988; Williams, 1995d). The marine to terrestrial interval in the Exploits Subzone overstep sequence accumulated conformably above the deep-marine deposits of the Ashgill–Llandovery Badger Group and underlying within-plate or volcanosedimentary arc successions without any evidence of a regional angular unconformity (McKerrow and Cocks, 1978; Williams, 1979; Williams, 1995b).

Plate 34. *Near Emily Cove in Burnt Bay, gently-dipping, grey–green sandstones from the lower part of the Wigwam Formation have bedding planes covered with ripple marks (upper bed) and polygonal sand-filled mudcracks (lower bed).*

Fossil-bearing rocks of mid Paleozoic age are found in the Botwood and Badger groups of the Exploits Subzone overstep sequence (Williams, 1972; Williams *et al*., 1993). However, in the central part of Notre Dame Bay, Silurian terrestrial strata are restricted to the Botwood Group and Silurian marine strata are confined to the Badger Group. Locally, at least, the marine to terrestrial transition is preserved in neither of these groups.

In central Notre Dame Bay, the regional distribution of the Lawrenceton and Wigwam formations is much more restricted than that of the underlying formations of the Badger Group (Figure 3). Although the Lawrenceton and Wigwam formations are closely spatially associated in most lithotectonic sequences of the Botwood Group within the map area, they are observed to be juxtaposed against a variety of rock groups other than the Goldson Formation of the Badger Group or its equivalents (Williams *et al*., 1993; Hughes and O'Brien, 1994; Dickson *et al*., 1995). The Bot-

Plate 35. *Cross section of a steeply dipping crevasse in gently-dipping basalt of the Lawrenceton Formation near Wiseman Head. Coeval terrestrial sedimentation and volcanism is indicated by the fact that the bedding in red sandstone within the crevasse is generally parallel to that of the overlying sedimentary strata in the background, that the vesicular basalt flow degassed toward and chilled against the outer walls of the crevasse, and that sediment infill was baked in the narrow pink zone seen adjacent to the crevasse wall.*

wood Group may have had direct stratigraphic links with older marine units of the Exploits Subzone (Plate 36; *see below*).

Fault Controls on the Botwood Group Basin

In the study area, the terrestrial succession of the Botwood Group is isolated from surrounding rocks by major fault zones and does not preserve any primary external stratigraphic relationships. The bounding fault structures formed during episodes of ductile regional deformation and are typically strata-parallel curviplanar thrusts (*see* Structure of the Notre Dame and Exploits Subzones in Central Notre Dame Bay, page 65). These have been regionally offset by northeast-trending brittle faults, such as the Northern Arm Fault and the Reach Fault (Figures 3 and 17).

Some of the northeast- and northwest-trending structural features observed in terrestrial strata have been previously postulated to have reactivated syndepositional structures near the arcuate margins of the Botwood Group (Kusky *et al*., 1987). These authors reasoned that the partially consolidated Badger Group underwent a protracted history of dextral shear during a soft collision with the rocks of the Notre Dame Subzone. Dextral transcurrent and allied normal faulting began with the opening of terrestrial pull-apart basins filled by the Early–Late Silurian Botwood Group, continued with regional folding and thrust faulting of the Lawrenceton and Wigwam formations, and ended with posttectonic transcurrent faulting along the Devono-Carboniferous Northern Arm fault system.

The present configuration of regional structures in the Botwood Group in Burnt Bay may reflect some original features of the basin architecture. Of particular importance is the imbrication of thin thrust sheets of the Lawrenceton and Wigwam formations on islands south of the Dunnage Melange and north of the main part of the Botwood Group (Figure 18). Within these discontinuous thrust sheets, the Lawrenceton Formation consists of vesicular and autobrecciated, purplish grey and green basalts overlain by felsic flows and felsic tuffs. The thin tracts of the Wigwam Formation also contain rocks regionally representative of the lower and upper parts of that formation. Significantly, the grey sandstone lithofacies of the lower Wigwam Formation locally contains conspicuous detrital clasts of black shale (Plate 36).

The nearly complete internal lithostratigraphy of the imbricate thrust sheets near Burnt Bay is important for three reasons. First, their regional geological setting is very similar to mineralized Botwood Group rocks in certain eastern Dunnage Zone gold prospects (Evans, 1993, 1996). In both areas, the Wigwam Formation of the Botwood Group lies close to quartz-feldspar porphyry and gabbro-infiltrated ductile fault zones (Dog Bay and Little Burnt Bay faults) that separate highly sheared olistostromal melange complexes (Duder Melange and Dunnage Melange) from thrust-bounded panels of the Badger Group (Sansom greywacke and Campbellton greywacke).

Second, the apparently 'allochthonous' thrust sheets of the Botwood Group may represent original outliers of a relatively condensed succession that lay above the Dunnage Melange, the New Bay Formation and the Badger Group, but which have subsequently become structurally detached from their depositional substrate. Such relatively thin outliers may have formed during a southwestward offlap of the terrestrial overstep sequence and developed above a northwest-trending fault block situated to the northeast of the main basin depocentre. Alternatively, the offlapped feature might have been an arcuate intra-basin uplift which contained a black shale-bearing map unit.

Plate 36. *Cross-section of an orange-weathered, light-grey sandstone from the lower Wigwam Formation on Rice Island in Burnt Bay illustrating abundant fragments of black graphitic slate with a strong preferred orientation. Though possibly an intraformational mud-chip conglomerate, the slate clasts are similar to structurally adjacent rocks within the Dunnage Melange and the New Bay Formation.*

Third, the southwestward paleoflow, which was dominant during deposition of the lower grey sandstone facies of the Wigwam Formation (Dec *et al*., 1993), implies that detritus was moving axially along the depositional basin for a period of time after the cessation of the Lawrenceton episode of volcanism. Thus, near the present northeastern margin of the Lawrenceton Formation, there appears to have been an uplifted region which possibly formed a depositional arch between the Botwood area and the Change Islands area to the northeast. The margin of the uplift may have been coincident with a Silurian growth fault, which was reactivated when the Botwood 'subbasin' became partially inverted during regional deformation.

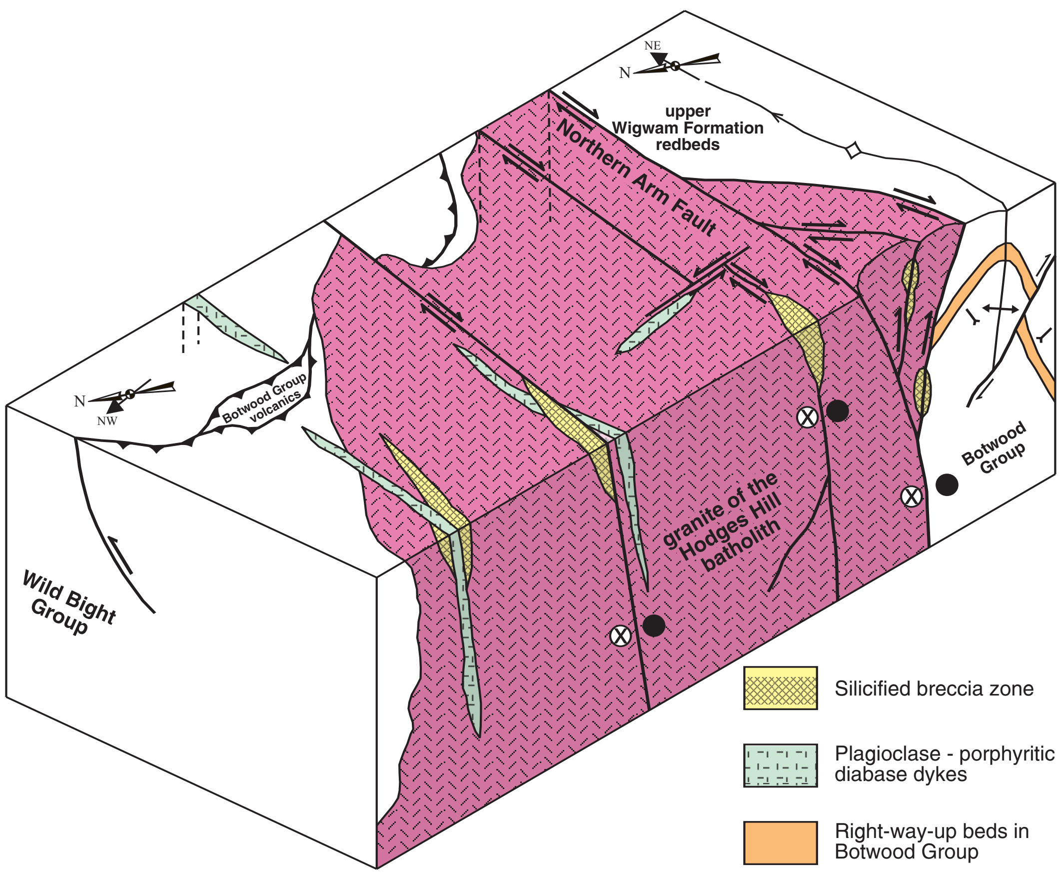


Figure 17. *Schematic block diagram of the Hodges Hill batholith in the vicinity of the dextrally transcurrent Northern Arm Fault. Note the association of plagioclase–porphyritic diabase dykes and silicified breccia zones with northeast- and northwest-trending strike-slip shears and oblique-slip thrusts. Note that some fold and thrust structures in Botwood Group country rocks predate batholith emplacement, whereas other brittle fractures are synchronous with those observed in the felsic plutonic rocks.*

Figure 18. *Detailed geological map of altered and mineralized rocks in the Botwood Group, where the unit is thrust imbricated with the Badger and Exploits groups. Second generation reverse faults are not displayed. Some mineralized Loon Bay-type felsic porphyries are also indicated. Locations of all known mineral occurrences are illustrated.*

STRUCTURE OF THE NOTRE DAME AND EXPLOITS SUBZONES IN CENTRAL NOTRE DAME BAY

The structural history of the central Notre Dame Bay region is presented by outlining the relative timing of the regional deformation episodes, by giving specific examples of the different types and scales of structures produced, and by highlighting the spatial variations seen in the development of the main phase structures.

Structures within the Notre Dame and Exploits subzones of north-central Newfoundland have been generally thought to reflect fundamentally different stages in the accretionary history of the Dunnage Zone (Table 2; Figure 5). In central Notre Dame Bay, previous workers have deemed that the Ordovician rocks of the Notre Dame Subzone display greenschist facies minor structures which predated terrestrial deposition of the Silurian Springdale Group (Nelson, 1981; Szybinski, 1995). These have been argued, on regional tectonic grounds (van Staal *et al*., 1990; Thurlow *et al*., 1992; Tremblay and Pinet, 1994; Cawood *et al*., 1995), as having been produced during the late Early and early Mid Ordovician (early and late Taconian), the latest Ordovician (Brunswickian) or the late Early Silurian (early Salinic). In contrast, mid Paleozoic tectonism in the Exploits Subzone has been commonly stated as being characterized by an initial Late Ordovician–Early Silurian soft-collision and melange-producing deformation (Pickering *et al*., 1988). This was then overprinted by a Late Silurian (late Salinic) and Early Devonian (early Acadian) hard-rock deformation which affected the terrestrial deposits of the Silurian Botwood Group (Lafrance and Williams, 1992; Figure 5).

In the area surveyed for this report, ductile structures of proven Ordovician age are restricted to the Exploits Subzone, although they are not regionally developed. Small folds, foliations and shear zones within parts of the South Lake Igneous Complex are Early Ordovician (O'Brien, 1992a; MacLachlan, 1998) and those observed within certain tectonically embrittled olistoliths of the Dunnage Melange are mid-Ordovician or older (Williams, 1992; O'Brien, 1993b; Lee and Williams, 1995). These local Ordovician features are generally coeval with regional tectonomagmatic events documented in Gander Zone psammites and Dunnage Zone ophiolites (Colman-Sadd *et al*., 1992) and southernmost Dunnage Zone felsic volcanic rocks (Tucker *et al*.,1994), although they may not be dynamically related to such events.

Folds, foliations and ductile faults of presumed Siluro-Devonian age are widespread in rock units of the Notre Dame and Exploits subzones. Major structures can be mapped throughout most of the area on 1:50 000 scale and appear to be related to observable mesoscopic and microscopic scale structural features.

REGIONAL DEFORMATION OF THE DUNNAGE ZONE IN NOTRE DAME BAY AND OTHER PARTS OF NORTH-CENTRAL NEWFOUNDLAND

In Notre Dame Bay, the regional deformation and dynamothermal metamorphism of the Early and Middle Ordovician rocks of the Notre Dame Subzone is stated to have begun in the Ordovician in areas immediately north and west of the Red Indian Line (e.g., Kennedy and DeGrace, 1972). There, pre- and post-Silurian thrusts and allied southeast-directed fold nappes root at or east of the Baie Verte Line (Hibbard, 1983; Szybinski, 1995). Such tectonism is unlikely to be related to the initial late Early Ordovician emplacement of the Dunnage Zone allochthons above the Humber Zone carbonate belt–the hallmark of northwestward obduction within the external Taconides (Williams, 1995c). Instead, these southeastward-directed structures are thought to record Late Cambrian accretion of a peri-Laurentian continental arc to ancestral North America and later accretion of the older pericratonic arc to a mid Ordovician continental margin volcanic arc. Thus, the Notre Dame arc was composite prior to the Silurian and mainly formed during westward subduction beneath the Taconic hinterland (Cawood *et al*, 1994; Swinden *et al*. 1997; Kurth *et al*., 1998; Figure 5).

A "hard-rock" deformation of the Cambro-Ordovician volcanic arc rocks in the Notre Dame Subzone has been interpreted to have occurred in the Ashgill during "wet-sediment" deformation of the Upper Ordovician Point Leamington Formation (Helwig, 1970; Pickering, 1987; Blewett, 1991). This was also thought to be associated with a southeastward structural translation of the Notre Dame Subzone rocks. In particular, the main melange belts within the Boones Point and Sops Head complexes were interpreted to have formed in the early Ashgill and been initially sheared at that time (Nelson and Casey, 1979; Nelson, 1981). Throughout the Late Ordovician and Early Silurian, Taconian-accreted rocks within the Notre Dame Subzone have been postulated to have been thrust southward or southeastward over the Ashgill and Llandovery turbidites and older strata of the Exploits Subzone (e.g., Dean and Strong, 1977; Johnson *et al*., 1994; Figure 5). In Exploits Subzone strata southeast of the Red Indian Line on New World Island, fault movements may have continued intermittantly until the late Llandovery, when some faults breached Early Silurian seafloor muds to produce the Joeys Cove olistostromal melange. There, Reusch (1987) postulated that fault-related deformation imposed a tectonite foliation on parts of the

Llandeilo Cobbs Arm Limestone, the postulated source of the marble olistoliths in the late Llandovery Joeys Cove melange.

Recent Structural Perspectives of the Eastern Notre Dame Subzone

The best documented evidence for the initial southeastward translation of the eastern part of the Notre Dame Subzone (Figure 5) comes from the Topsails–Buchans–Red Indian Lake region of west-central Newfoundland. There, Whalen *et al*. (1987) indicated that regional thrust-related deformation began sometime after the eruption of latest Arenigian rhyolite (the 473 +3/-2 Ma Buchans River Formation of Dunning *et al*., 1987) and ceased sometime before the intrusion of Wenlockian granite (the 425 ± 4 Ma plutonic phase of the Topsails intrusive suite). More recently, Thurlow *et al*. (1992) showed that the Kens Brook Volcanics of presumed mid Silurian age unconformably overlay the Hungry Mountain Thrust in the northwestern part of the Buchans region. This pre-Topsails mylonitic structure bounds the Early and Middle Ordovician metamorphic and plutonic rocks of the Hungry Mountain Complex (Whalen 1989) and places these relatively high-grade Notre Dame Subzone rocks structurally above the low-grade Early– Middle Ordovician Buchans Group.

Red Indian Lake

The Buchans Group has been interpreted to occur in a fault duplex structure which originated by progressive thrusting and collapse of the footwall plate beneath the overriding Hungry Mountain Complex (Thurlow and Swanson, 1987; Calon and Green, 1987; Figure 19).The uppermost (roof) thrust of the Buchans duplex is the pre-Wenlockian Hungry Mountain Thrust. The underlying openly folded reverse faults form a southeast-directed imbricate fan and are each associated with scaly foliated zones of block-in-matrix melange. The lowermost (floor) thrust of the duplex, the Victoria River Delta Fault, was postulated to have originally formed as the sole thrust of the Buchans Group (Figure 19). According to McKenzie and Desnoyers (1992), tectonic movements in the footwall of this fault caused melange formation and ductile shearing in late Llanvirn–Llandeilo times as quartz-feldspar porphyry bodies were emplaced into the underlying thrust sheet. However, Carboniferous normal fault reactivation of the sole thrust brought Silurian terrestrial strata, which may have lain with angular unconformity above the underlying panel of Ordovician Healy Bay–Harbour Round siltstone, into fault contact with the Buchans Group (Figure 19).

Green Bay

Discrete Ordovician and Silurian episodes of regional deformation have been reported on the headlands and islands of Notre Dame Bay immediately west of the area surveyed. There, Szybinski (1988), Szybinski *et al*. (1992) and Szybinski (1995) argued that Silurian compressional events in the Notre Dame Subzone accompanied the deposition of the Springdale Group but postdated the Ashgill–Llandovery and earlier overthrusting reported at the Red Indian Line (Figure 20). In the basement to the Springdale Group, Salinic structures in Early and Middle Ordovician rocks cut across and reactivated structures in the low-grade part of the Taconian fold-and-thrust belt.

Szybinski (1988) interpreted the Cambro-Ordovician rocks of the Lushs Bight and Western Arm groups (partial equivalents of the south-facing Moretons Harbour Group) to occupy a thrust-bounded fold nappe whose uplift history began in the late Taconic and continued into the early Salinic orogeny. Structural translation was initially southwestward during the Ordovician but, in later Ordovician and Silurian times, it became dominantly southeastward (Figure 20). Whereas some of the southeastward-directed thrusts in the Western Arm and Roberts Arm groups were interpreted to have moved during the Llandovery–earliest Wenlock (Szybinski *et al*., 1992; Figure 20), synchronously with foredeep accumulation of 429 ± 2 Ma terrestrial volcanic rocks (Coyle, 1990), southeast-over-northwest thrusting and strike-slip faulting postdated deposition of the youngest redbed conglomerate and sandstone units of the Springdale Group.

The pre-Middle Silurian deformation of the Roberts Arm Group was also governed by northeast-trending thrust faults, which are interpreted to have originally been situated below the sub-Springdale erosional disconformity (Figure 20). However, they may not have been the sole cause of the uplift and exhumation of part of the Roberts Arm stratigraphy (Bostock, 1988), as some of the erosion and denudation recorded beneath the unconformity surface may have been coeval with Silurian tectonism elsewhere in the Springdale basin. Regardless, the development of these pre-Silurian thrusts is contrasted with the northeast-striking Taconic fault structures found within the hangingwall sequence of Lushs Bight and Western Arm rocks above the Lobster Cove–Chanceport thrust (Szybinski, 1995; Figure 20). In the adjacent footwall sequence of the Lobster Cove–Chanceport fault, where imbricate thrusts of the Roberts Arm Group dip in the opposite direction, the earliest phase of displacement has been reported to be associated with southeast-over-northwest (Kerr, 1996) or east–west (Thurlow, 1996) tectonic transport directions.

Other Areas

Between Notre Dame Bay and Red Indian Lake, mid Llandovery or earlier deformation of Early–Middle Ordovician strata is reported to be constrained by isotopically dated intrusions near the Lake Bond deposit (pre-435 ± 2 Ma in Roberts Arm Group; Hudson and Swinden, 1990) and the Hammer Down deposit (pre-437 ± 4 Ma in revised Western Arm Group; Ritcey, 1993; Ritcey *et al*., 1995). However, a widespread north-over-south oblique thrusting event, which postdated a 429 ± 3 Ma pluton, has been reported to have affected rocks in the Topsails intrusive suite and overprinted

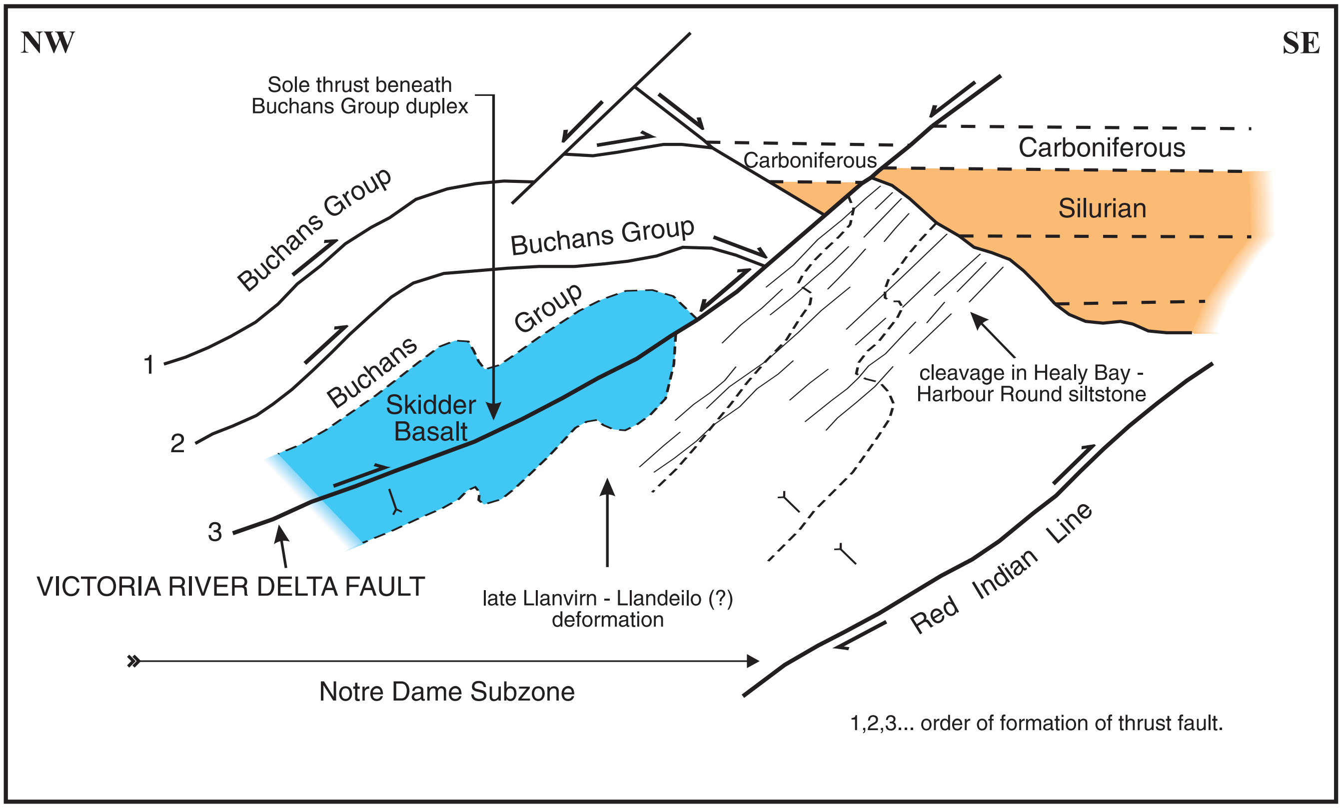


Figure 19. *Cross section schematically illustrating some of the main tectonic points of the Thurlow* et al. *(1992) interpretation of the structural evolution of the Buchans–Red Indian Lake area.*

earlier structures in that part of the Notre Dame Subzone (Whalen *et al*., 1993).

In the eastern headlands and islands of Notre Dame Bay, the age of thrust imbrication of the northwest and the southeast margins of the Cottrells Cove–Chanceport belt of the Notre Dame Subzone is partially known. Purported pre-Middle Silurian displacement on the northwestern boundary thrust (the Chanceport Fault) is interpreted to have initially been southeastward (Dean and Strong, 1977) or southward (Lafrance and Williams, 1992). Williams *et al*. (1976) and Williams (1995c) concluded that some rock units in the hangingwall sequence (the mid Cambrian Twillingate Granite and older Lushs Bight-equivalent rocks in the Sleepy Cove Group; Table 1) were mylonitized in ductile shear zones in the early Llanvirn. However, there is no evidence that the Moretons Harbour Group or the Chanceport Group (possible equivalent to parts of the Western Arm–Cutwell and Cottrells Cove groups) were regionally deformed at this time (Swinden, 1996). Moreover, it is unknown how these Ordovician structures relate, if at all, to the late Taconic fold-and-thrust belt in western Notre Dame Bay (Figure 20).

The Lukes Arm Fault, the southeastern boundary fault of the Chanceport–Cottrells Cove– Roberts Arm belt, is thought to have had a history similar to the Chanceport Fault (Lafrance 1989; Blewett, 1989), although deformation was dominated by late southeast-over-northwest, dextral oblique-slip thrusting (Figure 5). Significantly, the protracted movements described on the merged Chanceport and Lukes Arm fault zones (the Red Indian Line) have been argued to be entirely Late Silurian displacements (Elliot *et al*., 1991).

Recent Structural Perspectives of the Western Exploits Subzone

Throughout the Newfoundland Central Mobile Belt, previous workers have indicated that late Salinic deformation began with an episode of regional sinistral shear, which was controlled by the oblique convergence of the Exploits Subzone and the tectonic underplating of the western part of this subzone beneath the Notre Dame Subzone (Figure 21). Here, and in other parts of the northern Appalachians, Late Silurian deformation produced northeast-trending structures, such as cleavage-transected folds, oblique-slip thrusts and transcurrent mylonite zones with vertical and horizontal stretching lineations (Blewett and Pickering 1988; Currie and Piasecki, 1989; Holdsworth and O'Brien, 1993; O'Brien *et al*., 1993; Dube *et al*., 1992; van Staal *et al*., 1992; Dube *et al*., 1993; Lin *et al*., 1993; Lynch and Tremblay, 1992; Hibbard and Hall, 1993). The above-named features were overprinted by ductile structures produced in Late Silurian and Early Devonian episodes of dextral shear

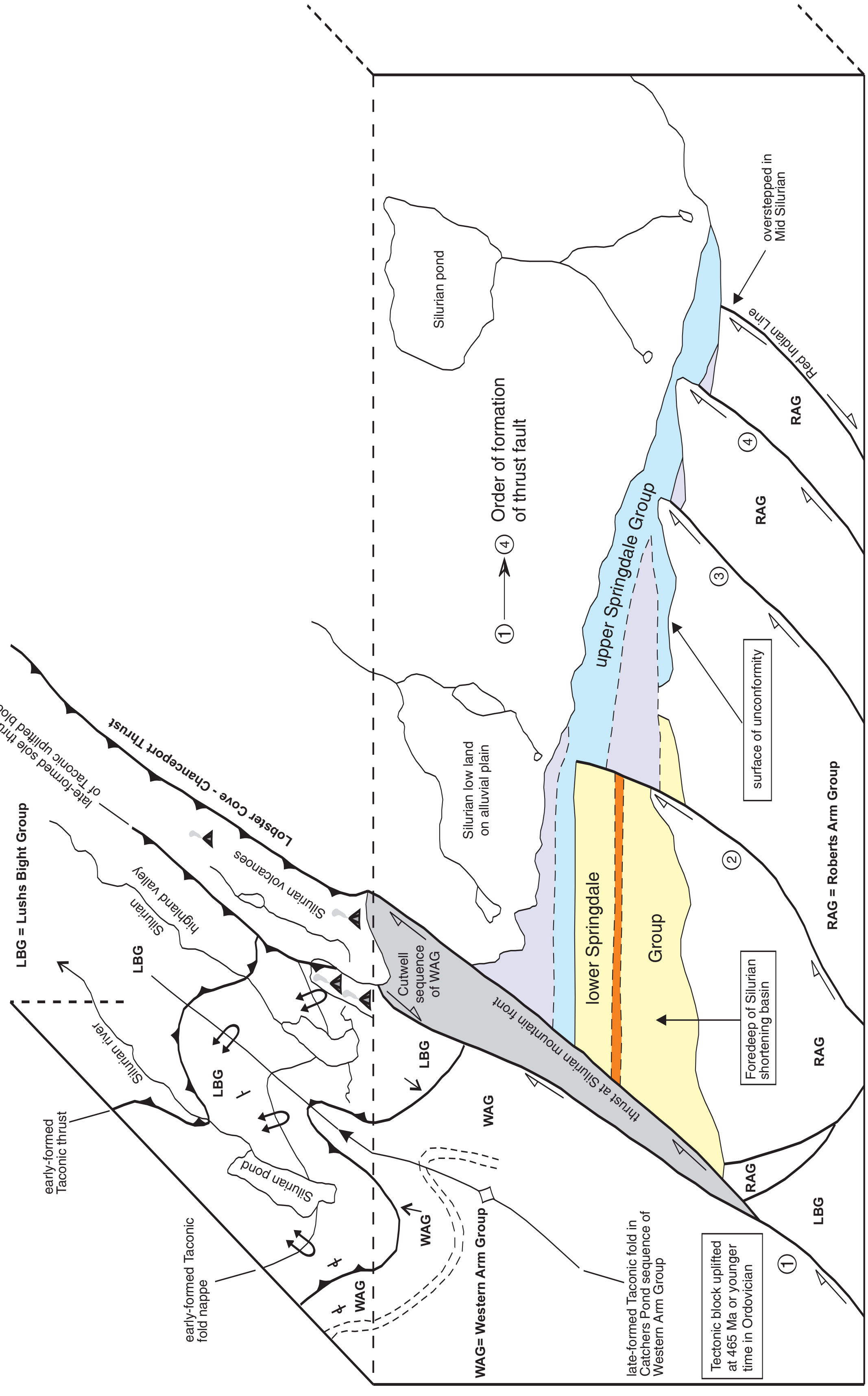


Figure 20. *Block diagram schematically illustrating some of the main tectonic points of the Szybinski (1995) interpretation of the Silurian evolution of the eastern part of the Notre Dame Subzone in western Notre Dame Bay.*

during the terminal collision of the Exploits and Notre Dame subzones (Figure 21).

In the eastern part of the Bay of Exploits, Lafrance and Williams (1992) confirmed Reusch's (1987) observations and stated that early southward-directed thrusting (Figure 21) probably produced the mylonitic foliation seen in Llandeilo marble clasts in the late Llandovery Joeys Cove Melange (part of the Badger Group). Although this compressional event was considered to have ceased prior to the deposition of the Middle–Late Silurian terrestrial strata in the Botwood Group, regional deformation associated with ductile dextral shear was thought to have affected this and older rock units in the Exploits and Notre Dame subzones (compare with Figure 20).

In central Notre Dame Bay, most ductile deformation of the stratified and plutonic rock units in the Exploits Subzone predated a 408 ± 2 Ma suite of minor felsic intrusions that were preferentially emplaced within and adjacent to the Loon Bay batholith (Elliot *et al.*, 1991; Figure 22a). Distinctive felsic–mafic composite dykes, which are possibly related to the Late Silurian (422 ± 2 Ma) suite on the Port Albert peninsula (Elliot *et al.*, 1991), are observable on the northern part of the Fortune Harbour peninsula. There, they crosscut folds and thrusts but are themselves locally sheared. Farther south, quartz-feldspar porphyry dykes, similar to those near the Loon Bay, Long Island and Northern Arm stocks, are present along the Ships Run of the Bay of Exploits, on the Little Burnt Bay peninsula and in the New Bay Pond area. These porphyries bracket local ca. 422 to ca. 408 Ma fault movements, as some dykes are openly folded or marginally foliated and others are undeformed (Figure 22a). A late suite of plagioclase porphyritic diabase and diorite dykes, which intrude certain felsic plutons of the Hodges Hill batholith and its country rocks (Dickson, 1998; Figure 17), provide a younger limit for regional ductile deformation.

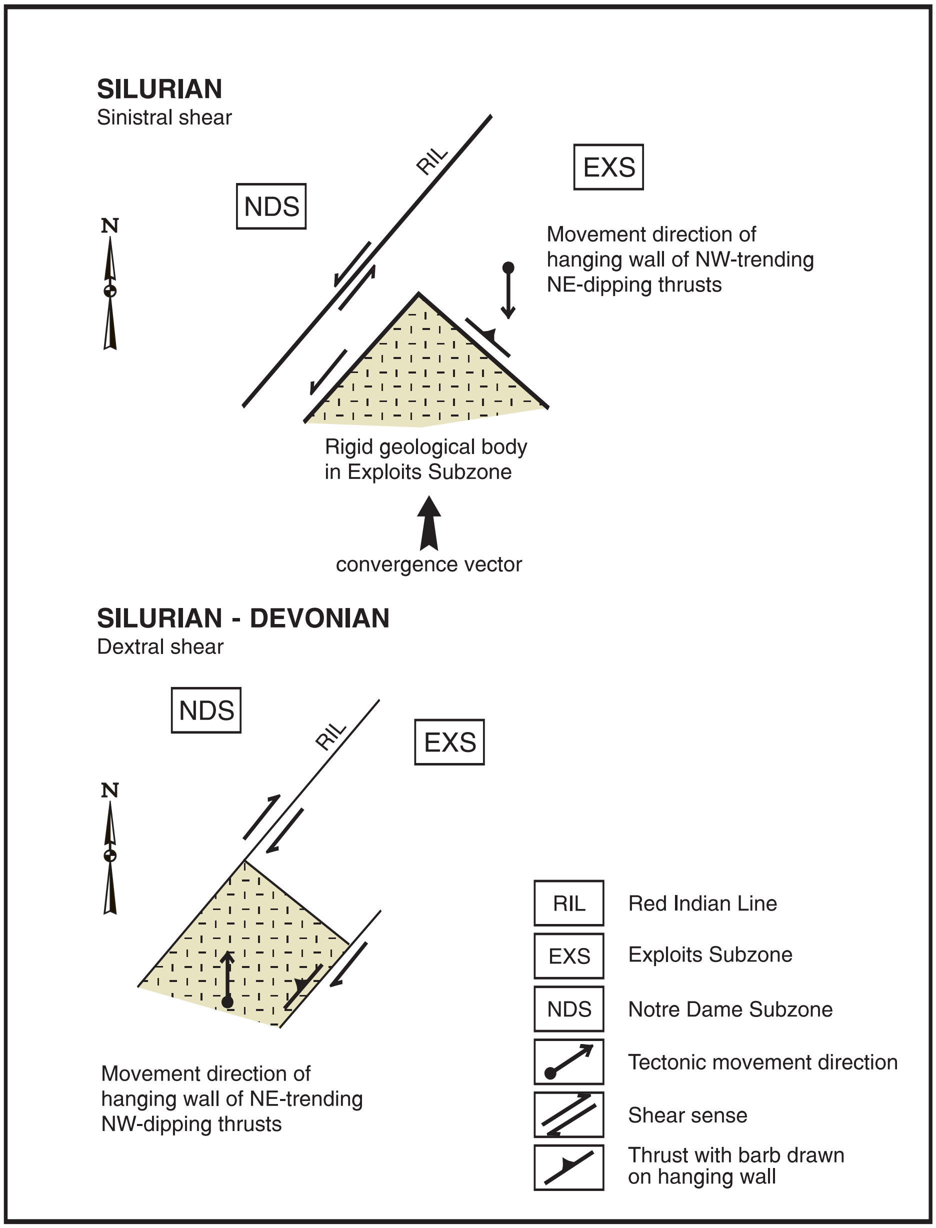


Figure 21. *Mid-Paleozoic tectonic map of the central Dunnage Zone illustrating how the movement direction of certain northwest- and northeast-trending oblique-slip thrusts may relate to the kinematic switch from sinistral to dextral shear along the collisional suture separating the Exploits and Notre Dame subzones (Red Indian Line).*

SILURO-DEVONIAN STRUCTURES

The central Notre Dame Bay area lies within a regional flexure in the Appalachian orogen outlined, in part, by segments of the Red Indian Line structural zone (Figure 3). Silurian structural features which define this Z-shaped oroclinal fold include regional slaty cleavage, fold axial surfaces, thrust faults, mylonite zones and melange belts. Brittle tran-

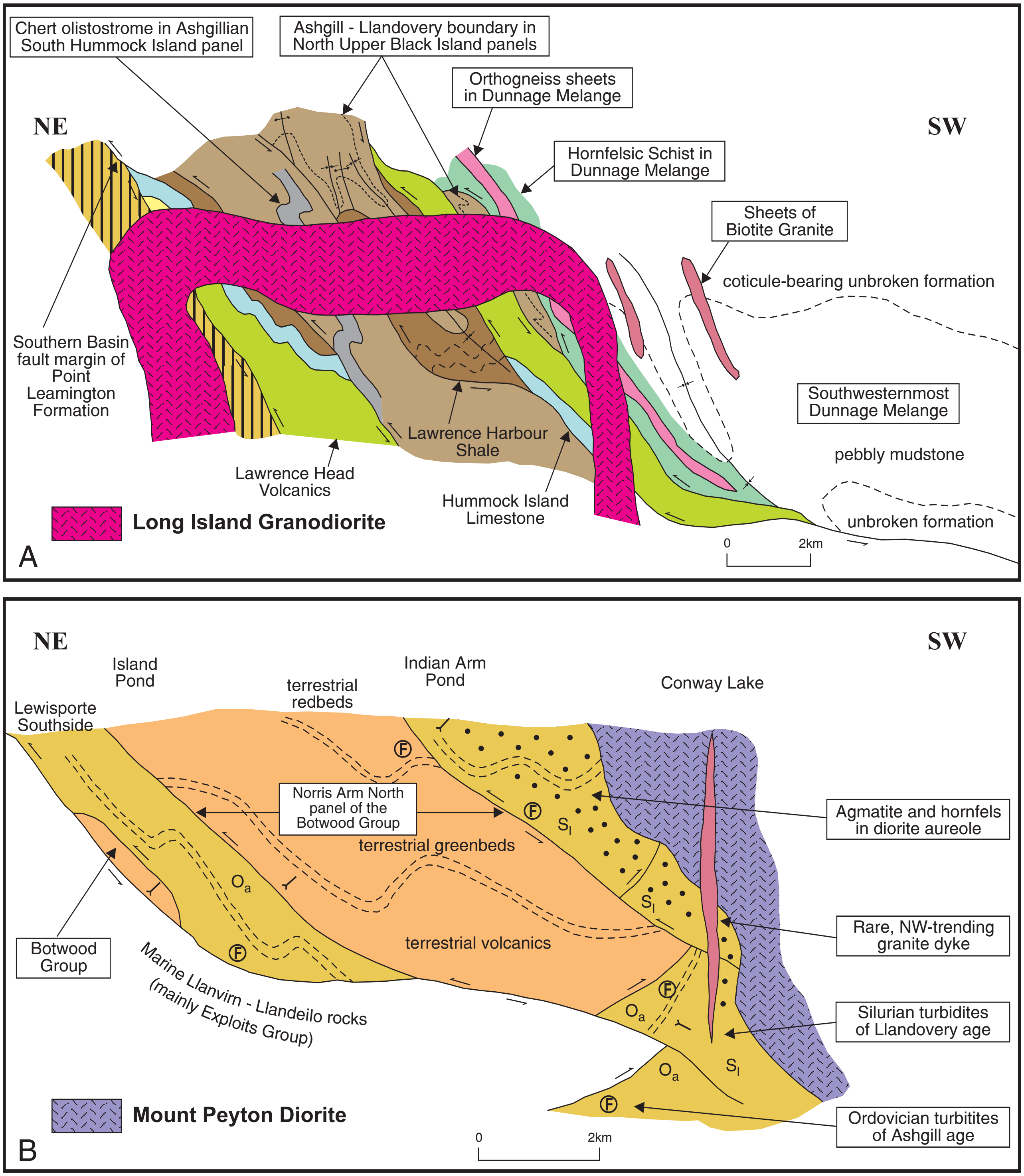


Figure 22. *Schematic cross sections emphasizing the nature of the southern margin of the Loon Bay batholith in the western Bay of Exploits and the northern margin of the Mount Peyton batholith south of Lewisporte. A) Detailed northeast–southwest section of the Long Island granodiorite and other sheeted intrusions within the fault zone separating the Point Leamington and other formations of the Badger Group from the Llanvirn–Llandeilo pebbly mudstone of the Dunnage Melange. B) Detailed northeast–southwest section of the Mount Peyton diorite and later granite dykes within the fault zone separating imbricated Badger and Botwood groups from the Llanvirn–Llandeilo rocks of the Exploits Group and the Dunnage Melange.*

scurrent faults of presumed Devono-Carboniferous age trend consistently northeastward and offset various segments of the Notre Dame Bay flexure.

In the area surveyed, ductile structures of regional extent are typically oriented northeastward, eastward or southeastward. This is regardless of the age of the rock unit affected or the Dunnage subzone to which it belongs. In all orientation domains, the structural translation of map units is commonly in two opposing directions, each sub-perpendicular to the local tectonic strike. Moreover, on the basis of most minor structures, reverse-shear displacement is apparently recorded for both movement directions (*see below*).

On lithospheric scale, the southeast- and northwest-dipping thrust faults that confront each other across the Red Indian Line (Quinlan *et al*., 1992) have been interpreted as conjugate reverse shear zones (Quinlan *et al*., 1993). These authors argued that the dominant northwest-dipping Silurian structures probably point to the structural underplating of the convergent Exploits Subzone and continued accretion of Paleozoic rocks beneath the Notre Dame Subzone (Figure 21). In contrast, the subordinate southeast-dipping Silurian structures probably reflect later tectonic readjustments which made Exploits Subzone rocks (and their late Precambrian basement) overplate the Taconian tectonites of the Notre Dame Subzone (Figure 5).

General Characteristics of Folds, Foliations and Faults

In many areas, gently plunging major folds control the disposition of volcanic and sedimentary sequences. Preserved fossils and original depositional features establish that many of these successions are in correct stratigraphical order. Typically, such strata occur within regional synclinoria or anticlinoria, where they are either unfoliated or have slaty cleavage at a high angle to bedding. It is common for folded strata to contain one (or more) foliations whose inclination varies from vertical (mostly in symmetrical folds) to gently dipping (mostly in asymmetrical folds).

In places, gently plunging conjugate folds have two steeply to moderately inclined axial surfaces that dip in opposite directions (Figure 23). Slaty cleavage is attendant, though not everywhere axial planar, to the inclined asymmetrical folds which define the box shape of the conjugate fold pair. In the map area, shape fabric lineations are rare but, where present, they indicate the ductile extension direction (i.e., the X axis of the strain ellipsoid) associated with the development of foliation in the hanging wall or footwall of a thrust fault. On the cleavage surface, extension lineations are observed to pitch either in the direction of maximum foliation dip or obliquely down-dip between the strike and the dip of the foliation (Figure 23).

As first proposed by Kay and Williams (1963) and regionally extended by Karlstrom *et al*. (1982), low-grade rocks in northeastern Newfoundland are structurally repeated by reverse faults. In many parts of the Exploits Subzone, these northeast-trending faults bound southeast-dipping panels of simply folded strata which, for the most part, are overturned and possess a moderately southeast-dipping slaty cleavage (Figures 23 and 24). However, both northwest- and southeast-younging stratigraphic successions have been reported in individual panels. These tectonic slices of right-way-up and inverted strata are commonly seen to alternate with one another and, taken together, they form an imbricated structural sequence of fossil-bearing rocks (Figure 24). Typically, each panel represents a single thrust sheet or a small fold nappe (e.g., Williams *et al*., 1988) whose bounding thrust faults cause omission or duplication of parts of the original Ordovician and Silurian stratigraphic column.

The magnitude of strain observed in folded and cleaved strata within a tectonic panel generally intensifies toward the bounding reverse faults; these regional strain gradients were established, for the most part, early in the history of Siluro-Devonian deformation. However, near major fault zones, the total strain recorded is heterogeneous and can be thus partitioned into a series of incremental strains, some of which were imparted by late phases of regional deformation.

In the area surveyed, chevron folds with inclined axial surfaces are angular-hinged and straight-limbed where the magnitude of strain is relatively low and the contrast in bed competancy of the well-layered strata is large. Aymmetrical folds have smaller wavelengths and display a strong S- or Z-sense of fold vergence in areas which record a larger strain. In poorly exposed areas of central Notre Dame Bay, one can recognize the presence of an unexposed ductile thrust by noting the variable nature of folds within regional strain gradients. Approaching major fault zones, shallowly plunging folds of a given generation may 1) become tighter and more asymmetric as they concomitantly increase their pitch on axial surfaces, 2) become strongly overturned toward the general direction of hanging wall transport (and have their axial surfaces become more gently inclined in the opposite direction), and 3) may ultimately be transformed into a fold that is neither an antiform nor a synform but is a sideways closing structure, whose fold axis lies parallel to the stretching lineation direction.

Where strata are ductilely attenuated and bed parallel-foliated, axial surface inclinations of minor folds of bedding planes commonly lie parallel to the hanging wall and footwall foliation and, in most cases, also to the intervening reverse fault. However, at some distance from the bounding faults of certain thrust sheets, the cleavage and fold axial surface have been observed to both lie at different angles to a bedding-parallel thrust (Figure 23). Where fault structures at the margin of a tectonic panel change from being steeply to moderately dipping, the regional slaty cleavage typically forms a primary half-fan in a vertical section in sympathy with the ductile fault (Figure 23). This suggests that the ductile thrust faults are simply the flatter components of listric reverse faults (Figure 24).

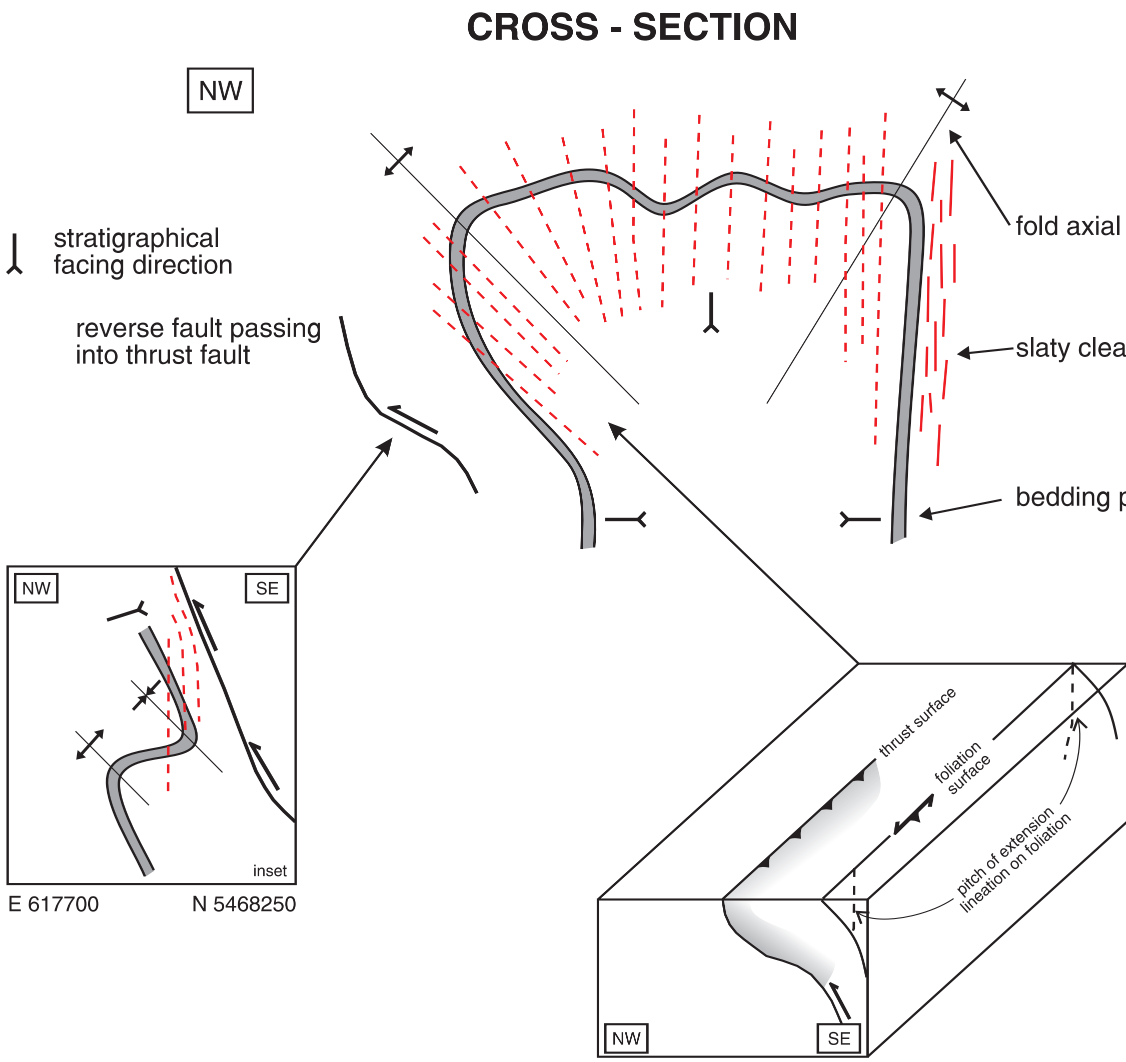


Figure 23. *Generalized northwest–southeast cross section of a very gently plunging conjugate anticline showing the variable relationships of regional slaty cleavage to bedding, reverse and thrust faults, and fold axial surfaces. Inset depicts fanning of non-axial planar cleavage near a thrust fault. Block diagram illustrates a common orientation of the extension lineation as seen on some moderately dipping slaty cleavage surfaces.*

Various parts of the strongly cleaved Lawrence Harbour Formation occur in imbricated thrust sheets and within thrust duplexes between Burnt Island in Osmonton Arm and New Bay's Little Northwest Arm (Figure 25). These structures are observable on metre and decimetre scale (e.g., E619350 N5479750) and can be mapped on kilometre scale along the well-exposed coastline. Furthermore, these lithotectonic sequences outline mesoscopic fold trains which are themselves truncated by thrust and reverse faults (Figures 25 and 26). Such tightly folded imbricate thrust faults develop where regional structures at opposing basin margins of the Point Leamington Formation converge southwest of the Boones Point Complex near the Red Indian Line (Figure 25 *inset*; O'Brien, 1990).

Foliated or folded rocks within early, middle and late Ordovician stratigraphic units pass laterally into block-in-matrix melange in the Red Indian Line thrust stack on islands in the Bay of Exploits (Plates 34, 36 and 37; Plate 6b in Lafrance, 1989). Farther southeast, well within the Exploits Subzone, tectonically straightened strata also develop narrow belts of tectonic melange in a major Siluro-Devonian fault zone at the boundary of the Dunnage Melange (Plate 8). Adjacent to these tectonic melange tracts, a greenschist facies slaty cleavage lies parallel to structurally- dismembered folds in proximity to small protomylonite zones (Plate 7).

SUPERIMPOSED REGIONAL DEFORMATIONS

Numerous studies documenting the detailed structural analysis of various parts of the Dunnage Zone in Notre

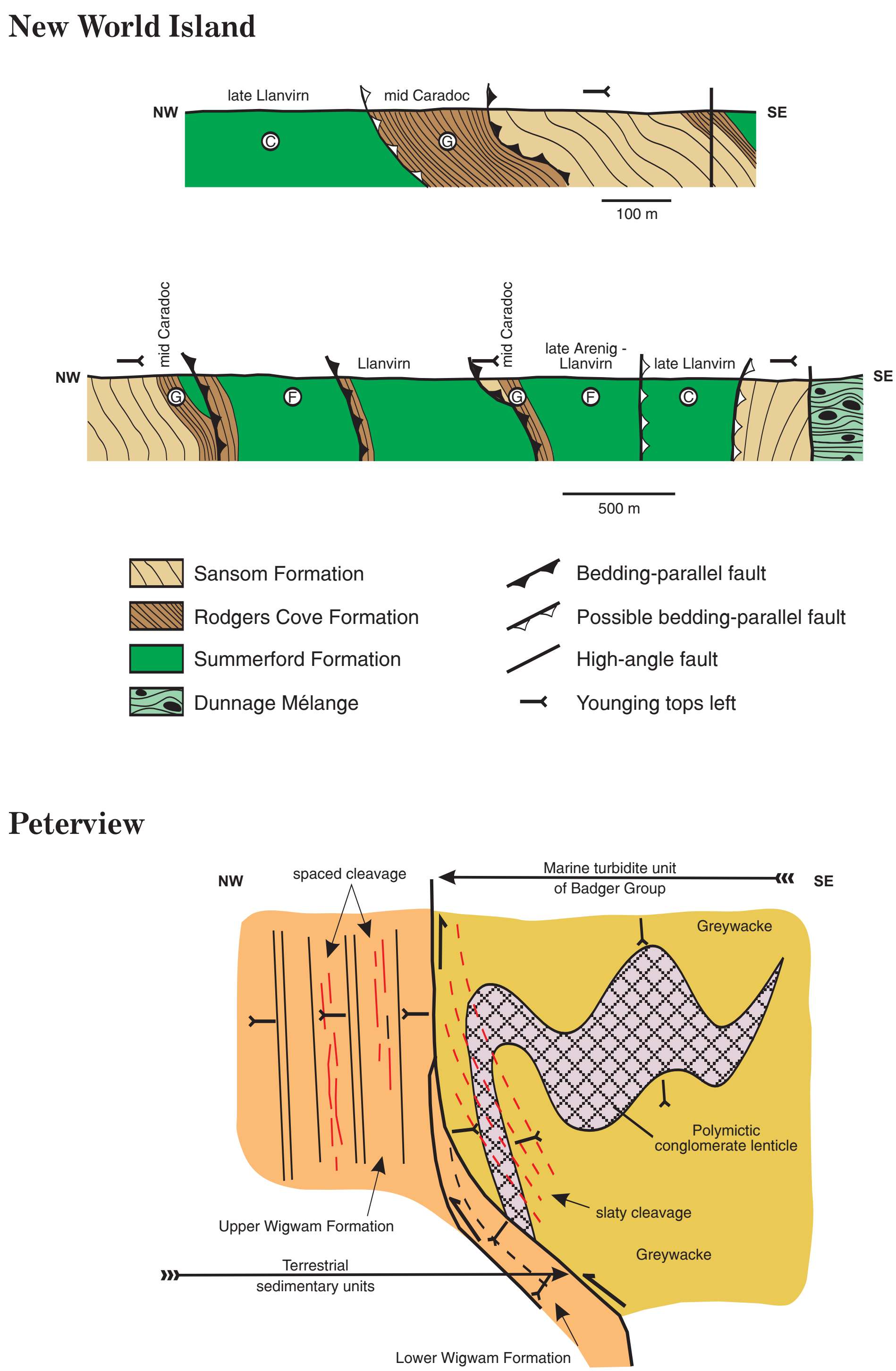


Figure 24. *Structural sections illustrating the disposition of fossiliferous Ordovician strata on part of New World Island (modified from Elliot* et al.*, 1989) compared to cross sections of Silurian strata near Peterview (modified from Dec* et al.*, 1993b). Note that northeast-trending, southeast-dipping thrust and reverse faults bound tectonic panels of simply folded and cleaved rocks. These faults separate northwest-younging inverted successions from southeast-younging right-way-up successions, and cause regional imbrication by omitting or duplicating parts of the Ordovician and Silurian stratigraphy.*

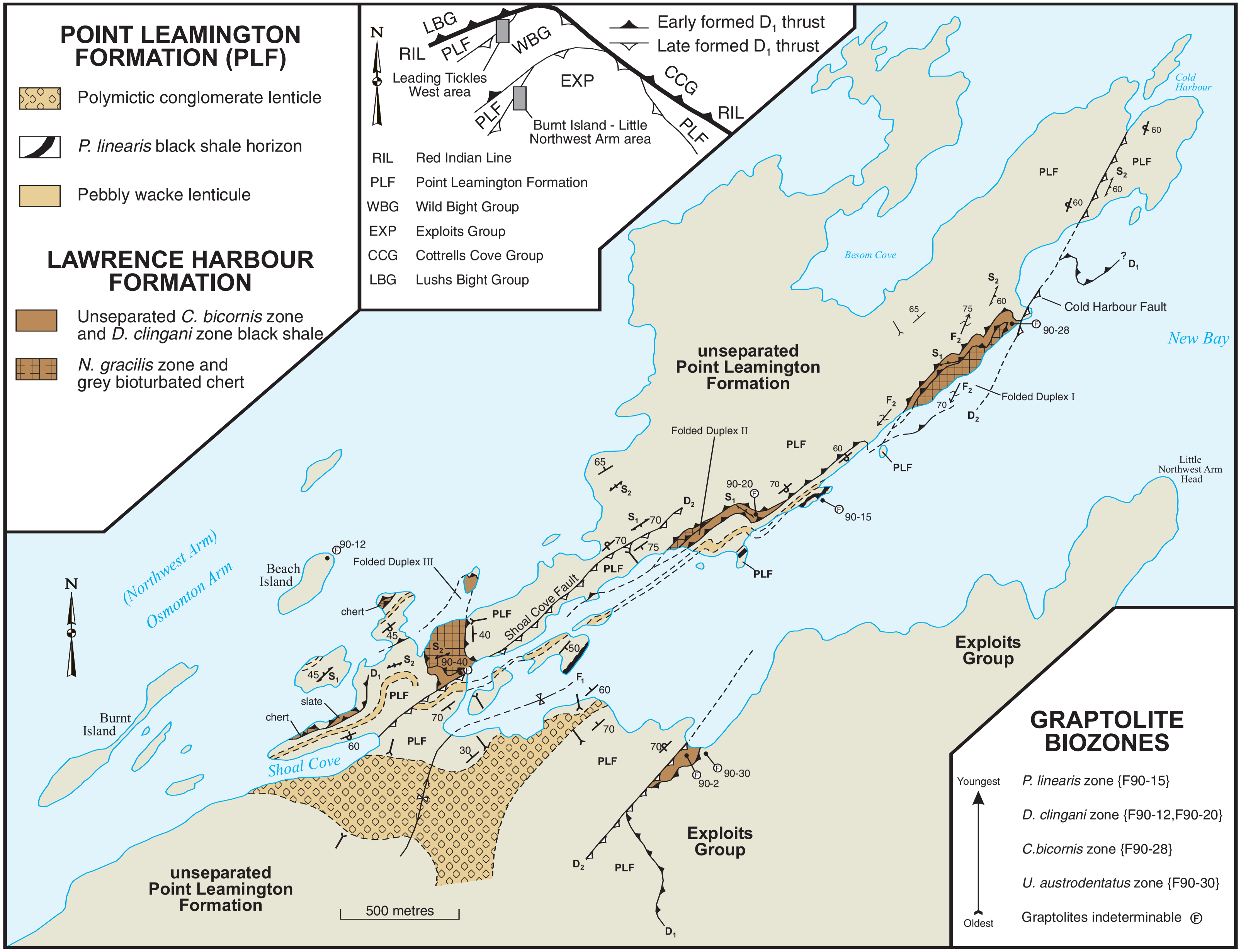


Figure 25. *Detailed lithological and structural map of three discrete, Z-folded thrust duplexes in the Burnt Island–Little Northwest Arm area. The duplexes are defined by various biostratigraphic units of the Lawrence Harbour Formation; the deformation also affects fossil-bearing Ashgill strata near the southeast margin of the Point Leamington Formation. Inset shows the tectonic location of this area relative to the D_1 thrust system of the Red Indian Line structural zone.*

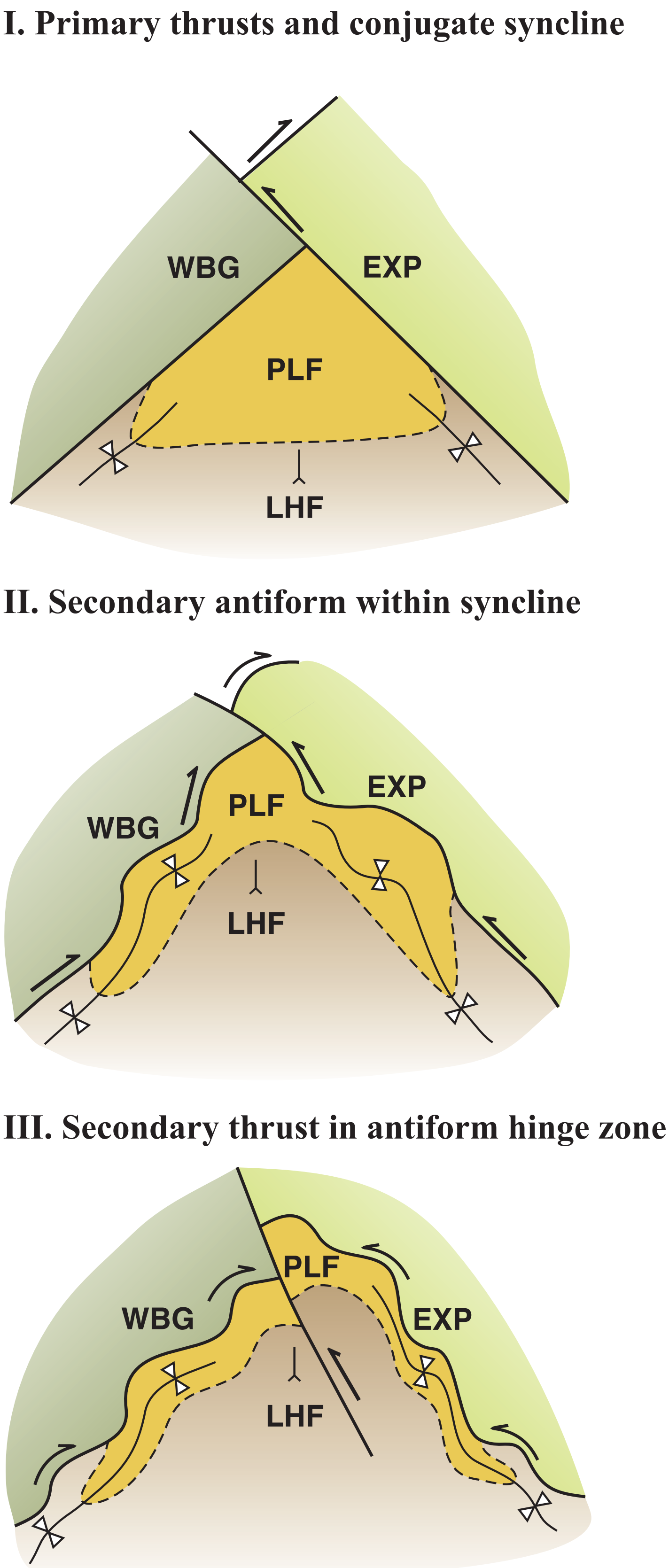


Figure 26. *Cross sections outlining a simplified portrayal of how the antiformally folded and re-thrusted duplexes in the Burnt Island–Little Northwest Arm area evolved from an original structural triangle zone. Rock unit abbreviations as for Figure 25. Phases (I), (II) and (III) represent a progressive temporal sequence of obliquely compressional events. WBG = Wild Bight Group, EXP = Exploits Group, LHF = Lawrence Harbour Formation, and PLF = Point Leamington Formation.*

Dame Bay have recently established complex sequences of overprinted tectonic structures which attest to superimposed polyphase deformation of Ordovician and Silurian strata (e.g., Nelson, 1981; Karlstrom *et al*., 1982; van der Pluijm, 1986; van der Pluijm *et al*., 1987; Blewett and Pickering, 1988; Elliott and Williams, 1988; Elliott *et al*., 1989). In the area surveyed, low-grade metamorphic structures were developed during four (D_1-D_4) phases of regional Siluro-Devonian deformation (O'Brien, 1993b).

D_1 Deformation

General Statement

In the centre of large D_1 tectonic panels, gently plunging F_1 open folds are commonly seen to have subvertical axial surfaces. In contrast, near the margins of D_1 tectonic panels, moderately plunging F_1 tight folds are mostly observed to be overturned toward the northeast or the southwest. Northwest-trending F_1 folds vary from gently to steeply plunging along the axial trace of regional anticlines and synclines. This is especially evident where such folds culminate or depress near the bounding D_1 faults of a particular map unit (e.g., E 613150 N5461550 adjacent to the Big Lake termination of the Point Leamington Formation).

Northwest-trending S_1 slaty cleavage is observed to dip vertically, northeastward and southwestward. It has highly variable and complex relationships with F_1 folds (*see below*). In places, L_1 extension lineations develop on S_1 cleavage adjacent to reverse faults and indicate that dextral oblique-slip movements occurred on some northeast-dipping D_1 Silurian fault structures (Figure 21).

Throughout the area surveyed, southwest-dipping D_1 reverse faults are those most commonly encountered; however, there are regions where northeast-dipping D_1 fault structures predominate or are equally well developed. For example, in the Northern Arm area of the Exploits Subzone (O'Brien, 1993a), the southwestern structural tract of the Exploits Group (O'Brien *et al*., 1997) was folded, translated southwestward and inter-

Plate 37. *Lawrence Head Formation basalt on southeastern Hummock Island passes into a zone of tectonic melange within a Red Indian Line thrust stack. Narrow, sharply bounded ultramylonite zone beneath hammer head contrasts with wider, diffusely bounded protoclastite zone below the end of the hammer shaft. Light-coloured "blocks" are distorted intrafolial lithons of the original basalt and the darker-coloured "matrix" has abundant neocrystallized chlorites derived from more comminuted parts of the protolith.*

nally imbricated prior to being overriden by the northeastward-directed imbricate thrust stack at the structural base of the Botwood Group (e.g., Lewisporte area of O'Brien 1992b; also Figures 18 and 22b). Hence, northeast-directed D_1 tectonic slices of mainly right-way-up successions of the Botwood Group are imbricated with previously folded and faulted sections of the Exploits Group. Thrust sheets containing parts of the Lawrenceton and Wigwam formations of the Botwood Group are structurally interleaved with a tectonic stack of regionally south-facing Exploits Group formations in the eastern Burnt Bay area west of Indian Arm Brook and a tectonic stack of regionally north-facing Exploits Group formations in the Southern Passage of the Bay of Exploits west of Stanhope (Hughes and O'Brien, 1994).

Examples of D_1 Structures

Most regions dominated by northwest-trending structure and stratigraphy display well preserved D_1 structures which are variably overprinted. In the Exploits Subzone, major D_1 faults are particularly evident where Caradoc black shale, Llandeilo chert and Llanvirn pillow lava have been tectonically removed from the local Ordovician succession. Structural excision of this characteristic part of the Middle Ordovician stratigraphic column from the top of the footwall plate occurs along the D_1 thrust separating the Wild Bight and Exploits groups between Northern Arm and Point Leamington, the D_1 thrust separating the Dunnage Melange and the Exploits Group on southern Upper Black Island and at Stanhope, or the D_1 thrust separating the Botwood and Exploits groups near Browns Arm (O'Brien, 1992b; 1993a).

The best examples of major F_1 folds occur in well-bedded turbidite successions within low-strain tracts of the Point Leamington Formation, the Campbellton greywacke and the New Bay Formation of the Exploits Group. A regional F_1 syncline is present in the Western Arm–Osmonton Arm area, where it affects the Point Leamington Formation, Goldson conglomerate and Randels Cove conglomerate (O'Brien, 1991a; Williams *et al*., 1992b). Another major syncline can be mapped in the Exploits Group between the Charles Brook–Sunday Island area and the Winter Tickle–Indian Cove area, where the New Bay Formation, Lawrence Head Formation, Strong Island chert and Lawrence Harbour Formation are all folded in succession (O'Brien, 1991b). The most extensive F_1 fold in the area surveyed is the Tea Arm Anticline which can be traced throughout the northeastern structural tract of the Exploits Group (O'Brien *et al*., 1997).

In the Notre Dame Subzone, the Moretons Harbour and Cottrells Cove groups strike regionally northwestward (O'Brien, 1990) and are strongly affected by imbricate D_1 thrust systems and major F_1 folds (Figures 6 and 27). Strong southwest-dipping S_1 cleavage is present near the boundaries of several large tectonic panels, is especially well developed in mainland exposures of the southern part of the Cottrells Cove Group south of North Harbour, and is ubiquitous within the Red Indian Line structural zone where it is gradational to protomylonite. Immediately southwest of the Lukes Arm Fault (Figure 3) small bedding-parallel D_1 faults, strong S_1 slaty cleavage and minor F_1 folds with non-axial planar S_1 cleavage are best seen in the black shales of the Lawrence Harbour Formation (*see also* Figure 25 where the D_1 structures in graptolitic black shales of the Exploits Subzone overstep sequence had predominantly dipped northeastward).

F_1 Periclines of the Exploits Subzone

Regional culminations and depressions occur along the axial traces of most major F_1 anticlines and synclines. The distribution of the various members and formations of the Exploits Group shows this feature particularly well (O'Brien, 1991b; Hughes and O'Brien, 1994). Periclinal

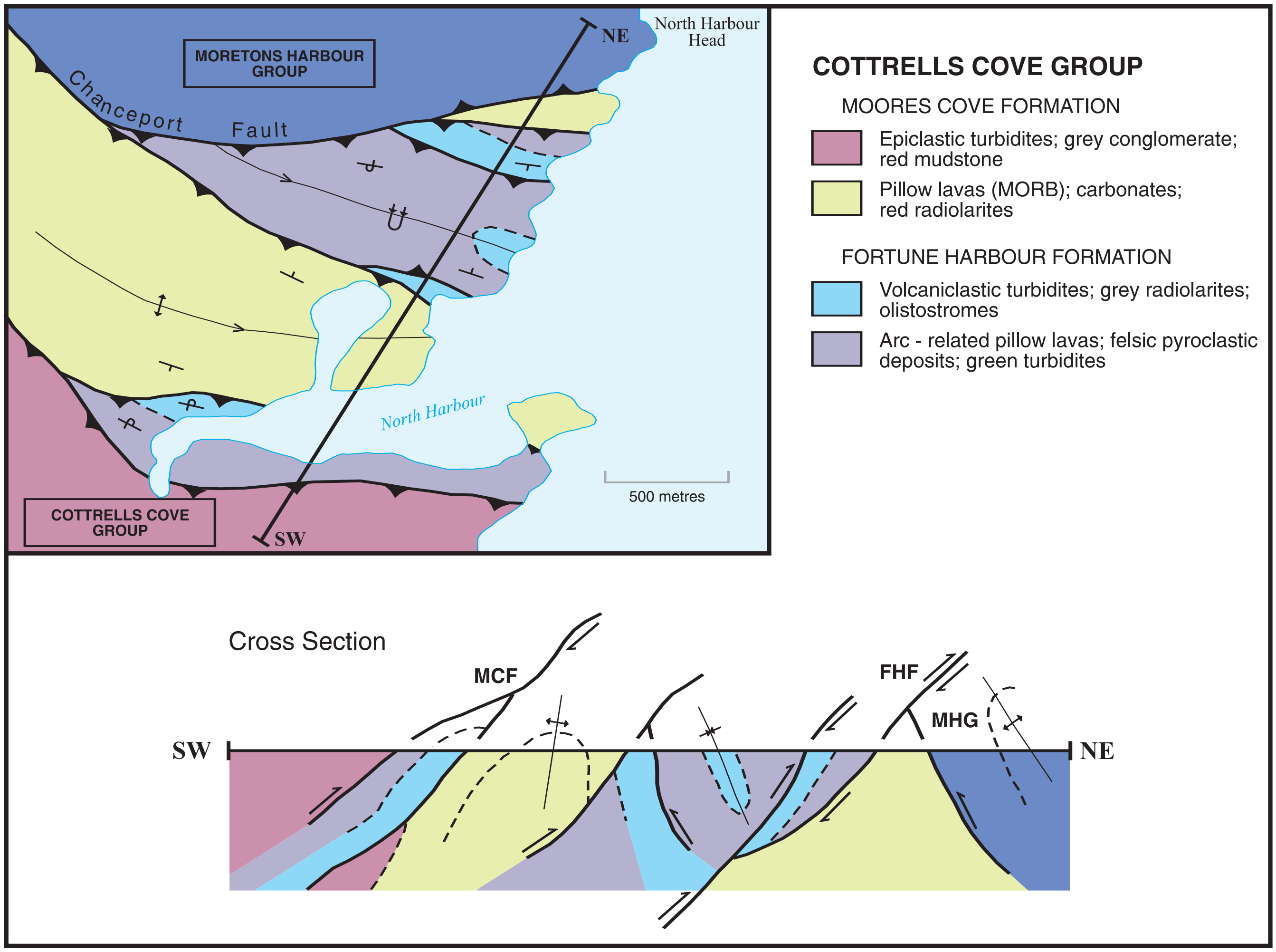


Figure 27. *Structural map of subunits of the Fortune Harbour and Moores Cove formations of the Cottrells Cove Group in the North Harbour area. Modified from Dec* et al. *(1997). Cross section illustrates an interpretation of the D_1 relationship of northeast- and southwest-dipping reverse faults and allied folds along this part of the east coast of the Fortune Harbour Peninsula.*

folds outlined by relatively resistent or magnetic rocks are observable on 1:10 000 scale aerial photographs and vertical gradiometer maps; large F_1 periclines can be accurately mapped at 1:50 000 scale (O'Brien, 1992b; 1993a and b) and readily compiled at 1:100 000 or smaller scale (Williams *et al*., 1992; O'Brien *et al*., 1997).

Where northwest-trending F_1 folds are refolded by northeast-trending F_2 folds, fold interference patterns are locally produced in negligibly strained rocks. Excellent examples of F_1 folds and S_1 cleavage overprinted by D_2 structures are seen in the New Bay Formation of the Exploits Group near Seal Rocks on Thwart Island, at the largest cove south of the lighthouse on Upper Black Island, and between Glovers Point and Indian Cove on the isthmus of the Fortune Harbour peninsula. The major F_1 periclinal syncline in the Point Leamington Formation is depressed within the outcrop of the Randels Cove conglomerate, which is the highest exposed stratigraphic subunit. However, this pericline is tightly refolded about northeast-trending axes to form a regional Z-shaped F_2 fold, which affects the entire depositional basin between Southwest Arm and Osmonton Arm (Williams *et al*., 1992).

Relationships of F_1 folds, S_1 Cleavage and D_1 Faults

Near the major F_1 Sunday Island Syncline (Helwig, 1967; O'Brien, 1990), southwest-dipping S_1 slaty cleavage, a gently doubly-plunging F_1 mesoscopic anticline and a small southwest-dipping D_1 high-angle reverse fault are overprinted by steeply southeast-dipping S_2 slaty cleavage and northeast-plunging F_2 open folds (Figure 28). The S_1 cleavage is observed to transect the F_1 fold hinge in a clockwise sense and to lie parallel to the D_1 reverse fault. This suggests that the D_1 deformation near Glovers Point was sinistrally transpressive (Figure 28) and, therefore, that at least some of the southwest-dipping D_1 faults had both left-handed strike-slip and reverse dip-slip components of dis-

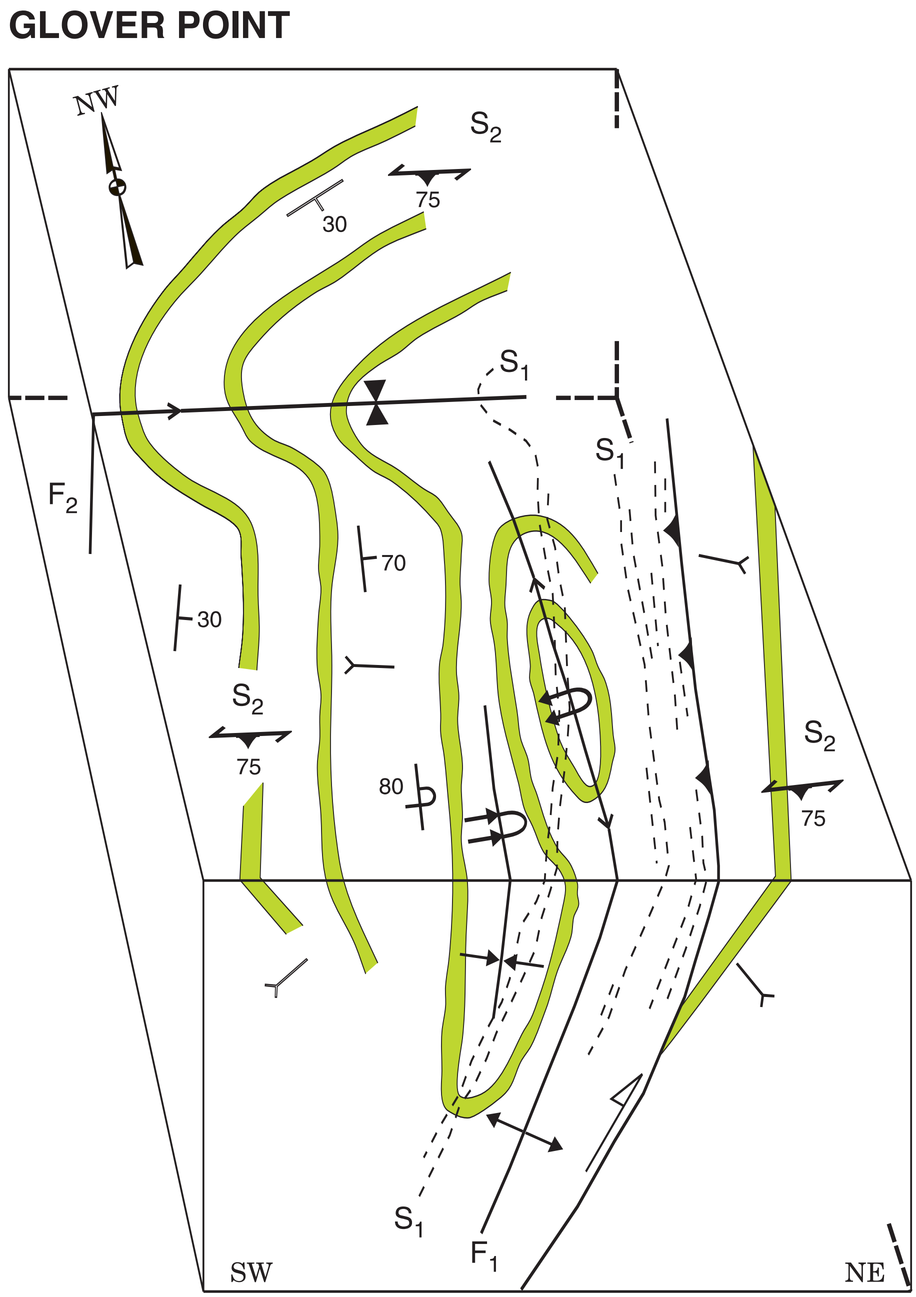


Figure 28. *F_1 folds transected by S_1 cleavage in the hanging wall of a southwest-dipping D_1 reverse fault near Glovers Point [E627900 N5464300]. Exposure is located in the New Bay Formation of the Exploits Group near the regional F_1 Sunday Island Syncline (Helwig, 1967). Note that northeast-trending D_2 structures overprint the clockwise-transected, doubly-plunging F_1 anticline.*

placement. The tectonic transport direction of the hanging wall plate of this oblique-slip structure was east-northeastward rather than up-the-dip of the reverse fault toward the northeast.

Coastal exposures of the New Bay Formation in the South Arm of New Bay include stratigraphically continuous turbidite beds on the low-strain limb of a gently plunging F_1 syncline. This succession of the Charles Brook member is situated some distance beneath a northeast-dipping D_1 overthrust located within the regional scale Paradise Fault Zone (Helwig, 1967; Figure 29). At exposure [E621650 N5465200], a southwesterly overturned F_1 anticline, which plunges both gently and steeply northwestward, is observed to be transected by S_1 slaty cleavage in an anticlockwise fashion (Figure 29). Moreover, the gently northeast-dipping S_1 cleavage is inclined at a shallower angle than the right-way-up beds. This indicates that D_1 deformation was dex-

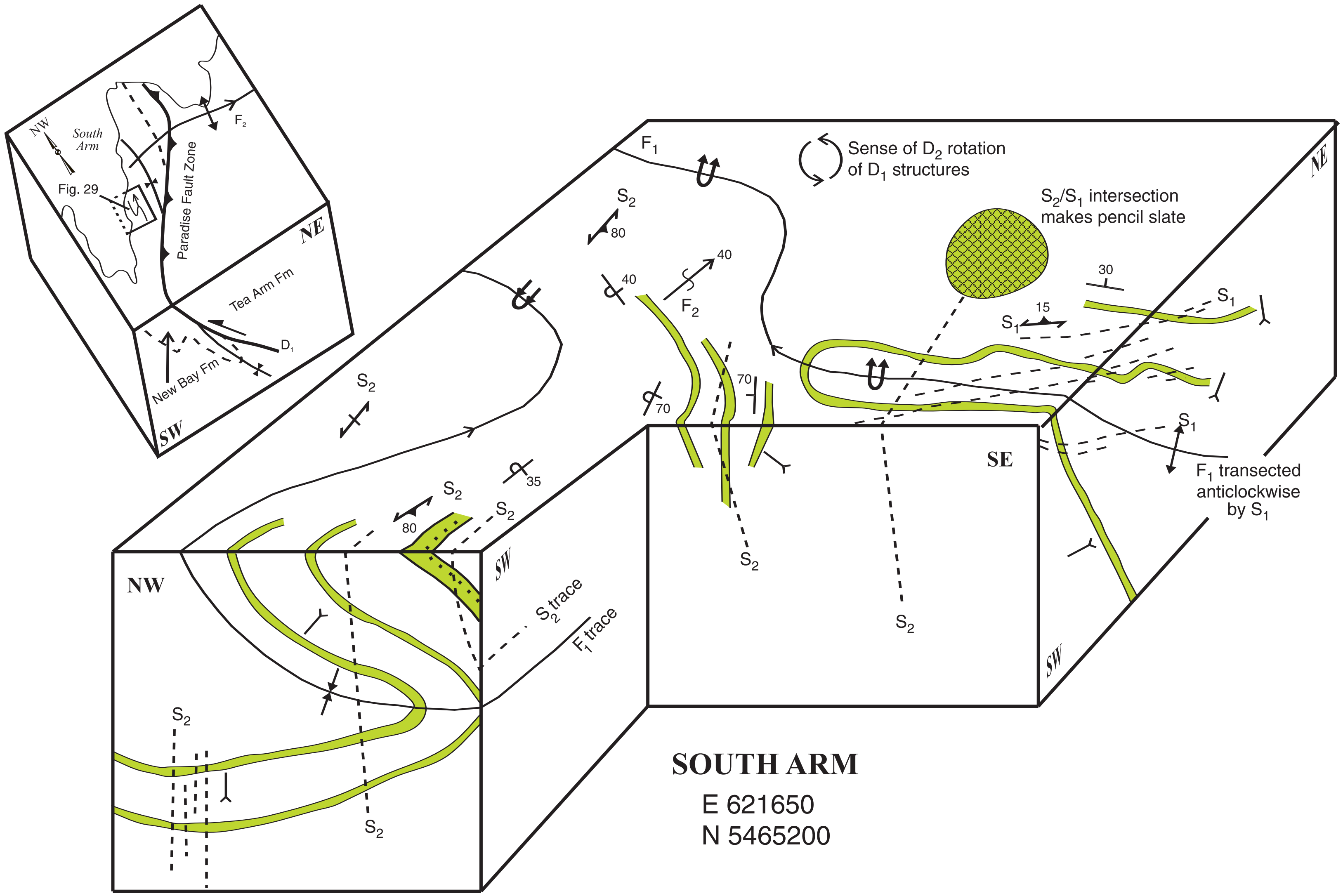


Figure 29. *F_1 folds transected by S_1 cleavage in the footwall of northeast-dipping D_1 thrust near South Arm [E621650 N5465200]. Exposure of S-shaped F_1 fold pair is located in the New Bay Formation of the Exploits Group on the right-way-up limb of a sheared-out syncline in the footwall sequence of the D_1 Paradise overthrust (inset). Note that northeast-trending D_2 structures overprint the anticlockwise-transected, northwest-plunging F_1 anticline. Minor northeast-trending D_3 structures observed at this locality were purposefully omitted.*

trally transpressive in this location and, therefore, that at least some of the northeast-dipping D_1 faults had both right-handed strike-slip and reverse dip-slip components of displacement. The tectonic transport direction associated with the dextral oblique-slip thrust near South Arm was between the south and the south-southeast.

In South Arm, the transected anticline is paired with a subrecumbent F_1 syncline in the footwall plate of the Paradise overthrust, and both folds were bodily rotated counterclockwise by large scale D_2 structures (Figure 29). As a result, northeast-trending S_2 slaty cleavage and minor F_2 folds overprint F_1 axial surfaces where the D_1 structures strike northeastward and northwestward. However, in many other places, it is difficult to distinguish the northeast-trending D_1 structures from the regional D_2 structures.

D_1 Reverse Faults of the Exploits Subzone

In the New Bay area, the Paradise Fault (*see* Map 2001-41 in pocket) is a moderately to steeply northeast-dipping reverse fault separating the Exploits Group into two D_1 structural tracts, one facing stratigraphically northeast and the other southwest (Figure 11). The northeast margin of the Exploits Group succession in the northeastern tract is a southwest-dipping D_1 thrust of the Red Indian Line structural zone (the Lukes Arm Fault). The southwest margin of the Exploits Group succession in the southwestern tract is a northeast-dipping D_1 reverse fault (the New Bay Fault; Figure 11). The Paradise and New Bay fault zones merge in the Strong Island Sound area to form a steeply northeast-dipping series of D_1 reverse faults. On the southwest end of Strong Island, the ca. 486 Ma pillowed arc-tholeiites of the Tea Arm Formation directly overplate the ca. 470 Ma pillowed alkali basalts of the Lawrence Head Formation. The Paradise and New Bay reverse faults also coalesce (at a branch point) near Phillips Head Pond near the tip of the thrust sheet which contains the Phillips Head igneous complex (O'Brien, 1993a; Figure 15). Thus, the Z-folded tectonic panel comprising most of the southwestern structural tract of the Exploits Group terminates along strike as a D_1 thrust slice to the northwest and the southeast (i.e., toward and away from the viewer of Figure 11).

South of Big Lake near the town of Point Leamington, the New Bay reverse fault merges with a northeast-dipping D_1 thrust belonging to the Northwest Arm fault zone (Figure 11). The southeastward termination of the type section of the Point Leamington Formation was caused by the coalescense of these basin-margin faults. Farther southeastward, such northeast-dipping D_1 faults placed the Exploits Group structurally above the Wild Bight Group (Figure 11). In places, the imbricated Exploits Group was also thrust above a narrow tectonic panel of the Shoal Arm Formation (Figures 3, 15 and 30).

Between New Bay Pond and Northern Arm, the Shoal Arm Formation is non-contiguous along its strike and contains folded D_1 faults which displace the Caradoc black shale and chert divisions of this formation (O'Brien, 1993a; Figure 3). Although a dominant stratigraphical facing direction is commonly recorded across the entirety of any one Shoal Arm section, controlled sampling of graptolite-bearing strata indicates that older *C. bicornis* Zone beds (F93-7) structurally overlie younger *D. clingani* Zone beds (F93-6) in some locations (Figure 30). This demonstrates that parts of the map unit are stratigraphically discontinuous, much like the tectonically adjacent Exploits and Wild Bight groups. Internal reorganization of the Shoal Arm Formation has been achieved by D_1 duplication and amalgamation of its primary constituent members, or through D_1 structural excision and fragmentation of parts of the unit.

Biostratigraphically constrained examples of D_1 thrust duplexes are found in the generally northwest-trending belts of the Shoal Arm Formation southeast of the town of Point Leamington (Williams and O'Brien, 1994). However, certain of these D_1 duplexes dip northeastward; whereas, others dip southwestward. Unlike the fossil-bearing D_2-reoriented D_1 duplexes of the Burnt Island–Little Northwest Arm area (Figure 25), D_1 imbricate thrust sheets near Northern Arm moved in two opposing directions to form small structural confrontation zones (Figure 30).

Northwest of Northern Arm, the Shoal Arm Formation illustrates a good example of the structural effects of the displacement of southwest-dipping D_1 reverse faults by northeast-dipping D_1 reverse faults. A southwest-dipping tectonic stack composed of right-way-up sedimentary and volcanic strata from the Pennys Brook Formation of the Wild Bight Group structurally overlies a thrust-bounded slice of right-way-up grey chert and Caradoc black shale from the Shoal Arm Formation (Figure 30). This lithotectonic sequence overplates a more stratigraphically complete section of the Shoal Arm Formation, whose youngest beds are right-way-up in the southwest part of the tract and whose oldest beds are inverted in the northeast part of the tract. The upside-down sequence of northeast-dipping red and grey chert from the lower Shoal Arm Formation is structurally overridden by a right-way-up, northeast-dipping succession of older Exploits Group rocks. These strata occur in the hanging wall of a northeast-dipping D_1 reverse fault and comprise parts of the Brooks Harbour and Saltwater Pond members of the New Bay Formation. Significantly, the graptolite localities within the black shales are located, structurally, within separate southwest-directed and northeast-directed thrust sheets. The Shoal Arm Formation's cross-sectional geometry is interpreted as a D_1 structural triangle zone in which black shale and chert are assumed to be completely detached from adjacent up-thrusted panels of the Wild Bight and Exploits groups (Figure 30).

A regional train of S-shaped F_2 folds (with attendant northeast-trending, vertical S_2 slaty cleavage) refold the D_1 thrust stacks near Northern Arm subsequent to the formation of the triangle zone. Where D_1 tectonic panels dip to the southwest (e.g., around ponds west of Route 350), the F_2 plunge is southwestward. Where D_1 tectonic panels dip to

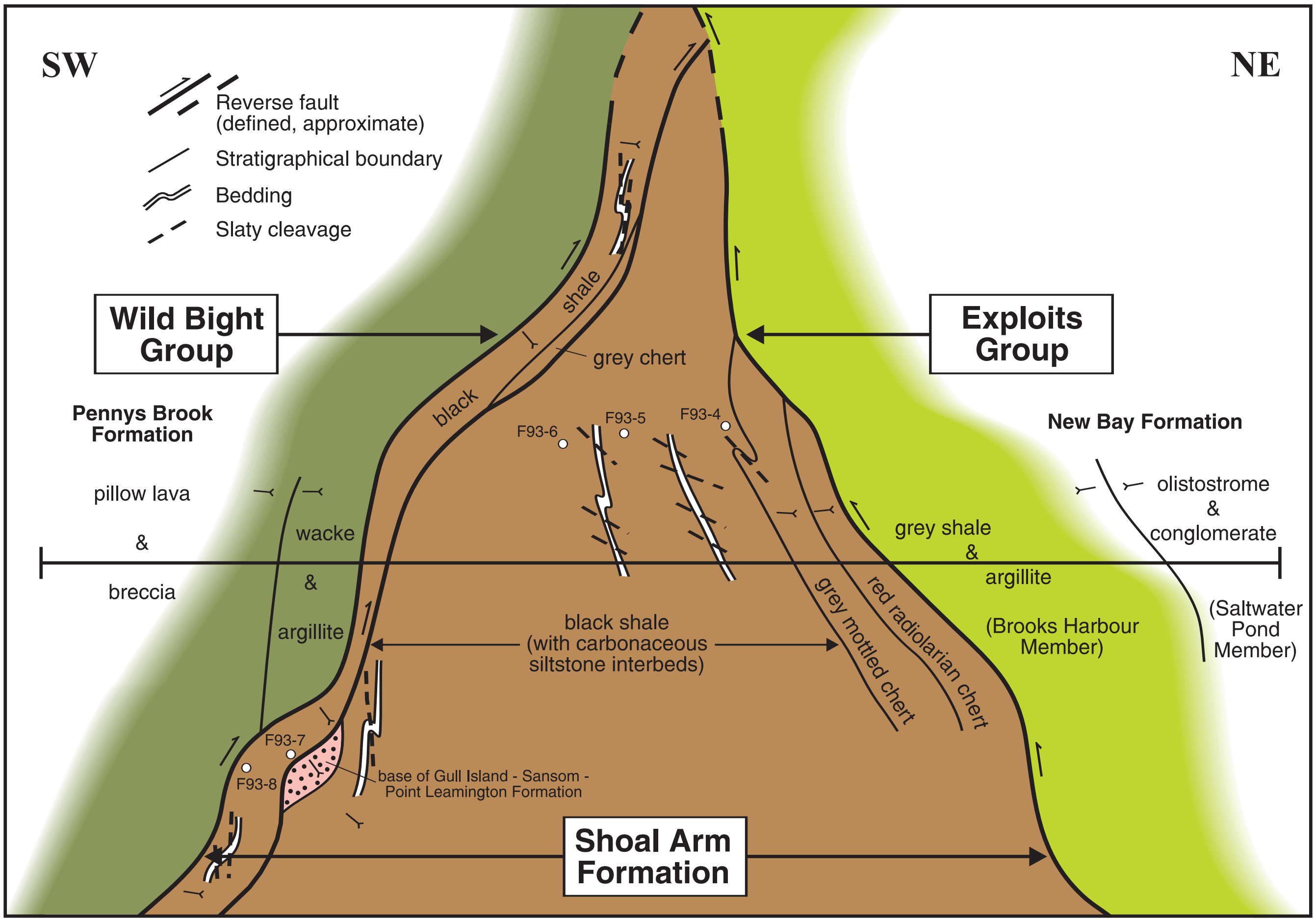


Figure 30. *Cross section, viewed looking northwest, showing the upper part of the structural triangle zone outlined by the Shoal Arm Formation near Northern Arm. The lithostratigraphic and structural position of the black shale-hosted graptolite localities is illustrated in each D_1 thrust sheet. Selected D_1 minor structures are schematically depicted; vertical scale is exaggerated. Modified from Williams and O'Brien (1994).*

the northeast (e.g., around ponds east of Route 350), the F_2 plunge is northeastward. Although similar to those depicted in Figure 26 (III), D_2 reverse faults in this region are minor and strike across the trend of the structural triangle zone.

D_1 Reverse Faults of the Notre Dame Subzone

From structural maps and sections of the Fortune Harbour and Moores Cove formations of the Cottrells Cove Group, drawn from their contact with the Moretons Harbour Group at North Harbour Head southwest to North Harbour (Figure 27), and also from Cottrells Cove southwest to their contact with the Exploits Group in Southeast Arm (Figures 7 and 8), it is evident that the southwest-directed reverse faults generally place older rock units over younger rock units. However, the northeast-directed reverse faults within and near this section (e.g., Plate 1 of O'Brien, 1993b) place younger rock units over older rock units in more locations than they put older rock units over younger rock units (Figures 7 and 27). The same is true of sections of the Sweeneys Island and Western Head formations of the Moretons Harbour Group, as drawn from North Head to Little North Harbour or from North Head to Indian Cove and farther southwest to the type area of Sweeneys Island (Figure 6). Common younger-over-older thrust contacts are predictable if the displaced lithotectonic sequence was already fault-imbricated. Thus, it is postulated that the early formed D_1 faults were southwest-directed and that the late formed D_1 faults were northeast-directed. Therefore, on regional considerations, it is most likely that the southwest-dipping faults displaced the northeast-dipping faults in this area.

The Chanceport Fault separates the Moretons Harbour Group from the Cottrells Cove and Chanceport groups in northern Notre Dame Bay and it is one of several regional structures that outline the Notre Dame Bay flexure (Figure 3). On the Fortune Harbour peninsula, on the middle limb of this Z-shaped oroclinal fold, the D_1 Chanceport Fault is marked, in most locations, by a steeply to moderately southwest-dipping reverse shear zone (O'Brien, 1990). In the

Cottrells Cove Group, thrust sheets containing northeasterly-overturned F_1 folds and/or having inverted, southwest-dipping stratigraphic sequences are observable at North Harbour (Figure 27) and farther west in the Southwest Arm of Fortune Harbour (on Gillespie Island and in Northwest Arm; Figures 7 and 10). Likewise, in the Moretons Harbour Group, an inverted southwest-dipping section through the Sweeney Island–Western Head transition zone is locally present in the hanging wall of the Indian Cove thrust (Figure 6).

Northeast-dipping D_1 belts of thrusts and folds are, however, also preserved in certain parts of the Cottrells Cove Group on the Fortune Harbour peninsula. A good example of such structures is present south of North Harbour Head (Figure 27). North of Gillespie Island (Figure 7), an undated quartz-feldspar porphyry intruding the Moretons Harbour Group outlines gently northwest-plunging, Z-shaped F_1 folds adjacent to thin mylonite bands in the Chanceport fault zone [E626650 N5486450]. This type of D_1 deformation may possibly relate to the overplating of the southwesterly-overturned anticline in the Sweeney Island Formation (Figure 6) above a footwall sequence of the Fortune Harbour Formation during dextral oblique-slip thrusting (northeasternmost reverse fault in Figure 27). The northwest-trending, northeast-dipping Silurian structure in Figure 21 may be analogous.

Large scale D_1 structures also face in opposing tectonic directions along the southwestern margin of the Cottrells Cove Group, where they form a major structural confrontation zone within and near the regional F_2 cross-folded Lukes Arm–Sops Head fault zone (Figure 7; Blewett, 1989). Near Moores Cove, on the northeast limb of the major Cottrells Cove Syncline, F_1 folds and D_1 thrusts are generally southwest-directed; whereas, on the southwest limb, F_1 folds and D_1 thrusts are dominantly northeast-directed. Northeast-trending F_2 open folds overprint both the northeast- and southwest-dipping D_1 reverse faults.

The Boones Point Complex is restricted to the southwestern flank of the Cottrells Cove Syncline. There, the complex is regionally fault-imbricated, most commonly with the Moores Cove Formation of the Cottrells Cove Group, but also with other formations of the Notre Dame and Exploits subzones (Figure 7). The margins of the tectonic panels that contain the various olistostromal melange tracts are generally highly sheared.

Along the northern flank of the Cottrells Cove Syncline, a northeast-dipping D_1 reverse fault juxtaposes the stratigraphically lowest parts of the Moores Cove Formation with the oldest subunit of calc-alkalic pillowed basalt in the Fortune Harbour Formation (Figure 7). Here, the thick subunit of volcaniclastic turbidites in the Fortune Harbour Formation, which is well developed near North Harbour and on the Duck Islands, is notably absent (compare Figure 27; O'Brien *et al*., 1994; Dec *et al*., 1997). In Josiah Spencer Cove (Figure 7), reverse shear zones related to the D_1 fault have strong S_1 foliation dipping 60° toward 015° and discernible L_1 shape-fabric lineations pitching 45° toward the northwest on the S_1 surface [E621700 N5485100]. Such extension lineations are consistent with dextral oblique-slip displacements on these northeast-dipping ductile fault structures. Moreover, they indicate that older Fortune Harbour basalts in the upper thrust plate moved toward the south-southeast above younger Moores Cove basalts and turbidites in the lower thrust plate (Figures 7 and 31).

A northeast–southwest cross section of the D_1 structural confrontation zone developed at the southwest margin of the Cottrells Cove Group illustrates how this tectonic feature is centred on the olistostromal melange tracts of the Boones Point Complex (Figure 31). A southwest-dipping imbricate thrust stack, over 750 m in structural thickness, contains the boundary between the Exploits and Notre Dame subzones. From its structural base to its top, this stack consists of tectonically straightened or inverted slices of the lower parts of the Moores Cove Formation (detached from the rest of the Cottrells Cove Group), a variably sheared lens of pebbly mudstone and block-in-matrix melange (the Southeast Arm panel of the Boones Point Complex), a thin S-folded thrust sheet of the Point Leamington Formation, and an overplated section from the middle and upper parts of the Exploits Group. The underlying northeast-dipping thrust stack is thought to contain mainly right-way-up imbricated fragments of the Fortune Harbour and lower Moores Cove formations (from the southern flank of the Cottrells Cove Syncline) thrust directly above the Yates Point panel of the Boones Point Complex.

The southwest-dipping overthrusts of late D_1 age that characterize the southern margin of the Lukes Arm–Sops Head fault zone probably also occur northeast of the Chanceport fault zone within the Moretons Harbour Group. Evidence comes, in part, from the complex D_1 relationships between folds and thrusts near the Indian Cove fault (Figure 6), a structure located in the Notre Dame Subzone about 12 km northeast across strike from the Lukes Arm–Sops Head fault zone. Regionally, the Western Head Formation of the Moretons Harbour Group becomes overfolded and inverted toward the northeast. In the hanging wall of the southwest-dipping D_1 thrust at Indian Cove, it faces stratigraphically toward a right-way-up footwall succession of the older Sweeney Island Formation (Figure 6). Along strike in the hanging wall sequence, 2 km farther west on the opposing headland of Fortune Harbour [E627250 N5488750], a moderately northeast-dipping thrust offsets the southwest-dipping inverted stratigraphic boundary of the Sweeney Island and Western Head formations (Figure 6, *inset*).

Thus, in this region, at least one northeast-dipping thrust fault is known to be younger than the northeasterly-overturned F_1 fold whose upside-down limb is situated in the immediate hanging wall of the southwest-dipping D_1 Indian Cove thrust (Figure 6 inset). The younger structure is interpreted as a minor D_1 back thrust affecting the hanging wall plate of the Indian Cove thrust fault; it indicates that

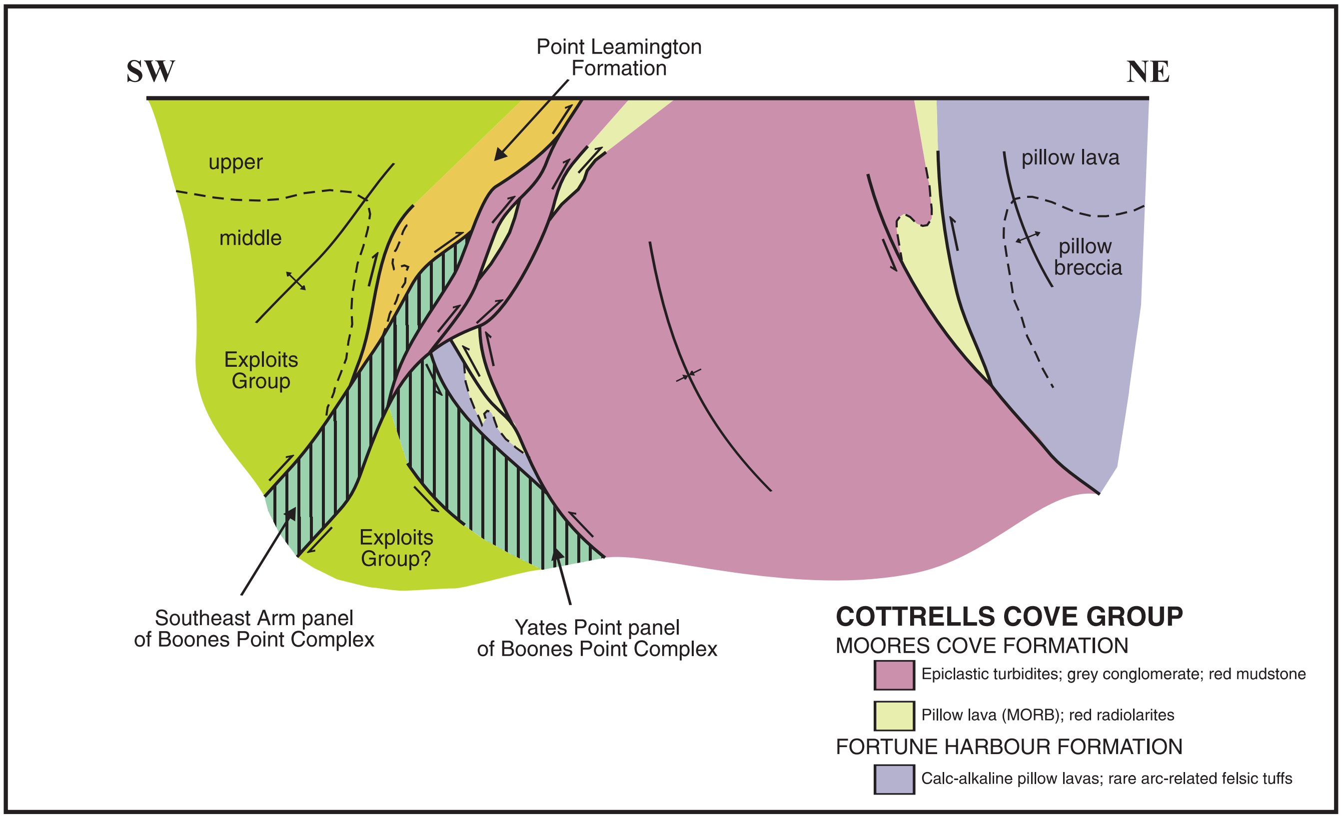


Figure 31. *Regional cross section, viewed looking northwest, highlighting the D_1 structural confrontation zone centred on the olistostromal melange tracts of the Boones Point Complex. A southwest-dipping imbricate thrust stack is interpreted to have been emplaced above a northeast-dipping imbricate thrust stack. The boundary between the Notre Dame and Exploits subzones is coincident with the highest imbricate thrust to affect the Boones Point Complex and the Moores Cove Formation.*

tectonic movements in certain late D_1 fault zones were probably protracted. On the basis of its orientation alone, this type of northeast-dipping structure can be easily confused with the early formed D_1 thrusts which predate the Indian Cove Fault (e.g., Figures 7, 27 and 31) and that are associated with regional southwesterly-overturned F_1 folds (e.g., the anticline cored by the Sweeney Island Formation in Figure 6).

Asymmetrical F_1 Folds of the Notre Dame Subzone

Along the linear arms of Fortune Harbour, in the type area of the Fortune Harbour Formation of the Cottrells Cove Group, the hinge zones of northeasterly-overturned F_1 asymmetrical folds are displaced by southwest-dipping D_1 thrust and reverse faults (Figure 10). In this particular location, the Chanceport Fault underlies these structures and also dips southwestward.

Where Z-shaped F_1 folds in one thrust sheet are juxtaposed with Z-shaped F_1 folds in an adjacent thrust sheet, the D_1 imbricate stack is dominantly composed of right-way-up, southwest-dipping rocks. The hanging wall and footwall sequence of the Gillespie Island thrust in Southeast Arm is representative of this type of stacked lithotectonic sequence (Figure 29). Thus, in the vicinity of the Cook Iron Mine, the structural sequence of Fortune Harbour Formation strata reflects a tectonic duplication of the primary stratigraphic successsion. Tectonic duplication of the Fortune Harbour Formation is more obvious in Northwest Arm, where fold trains dominated by northeasterly-overturned S-shaped asymmetrical folds are also observed to be tectonically juxtaposed against Z-shaped fold trains of Fortune Harbour basalt, tuff and chert. Near the Grey Copper Mine, the presence of megascopic S-shaped F_1 folds in the southwest-dipping imbricate stack is confirmed by the local preponderance of inverted stratigraphic sections that become younger toward the northeast.

In the vicinity of North Harbour on the Ships Run of the Bay of Exploits (Figure 27), a thick well-bedded succession of Fortune Harbour turbidites (Dec *et al*., 1997) overlies the same mafic and felsic volcanic rocks seen in the type area of the Fortune Harbour Formation farther west (Figure 10). Although the lower part of the succeeding Moores Cove Formation is also present, Moores Cove strata are everywhere in D_1 faulted contact with the volcanic and sedimentary units of the Fortune Harbour Formation. In the coastal section near North Harbour, southwest-plunging F_2 antiformal synclines and F_2 synformal anticlines occur in upside-down stratigraphic sequences of both the Fortune Harbour and Moores Cove formations. These D_2 structures overprinted the D_1 structures and developed on the long limbs of gently northwest-plunging, northeasterly-overturned, S-shaped F_1 asymmetrical folds.

On the west coast of the Fortune Harbour peninsula, F_1 asymmetrical folds displaying the opposing sense of D_1 tectonic polarity are present on both flanks of the regional Cottrells Cove Syncline (Figure 8). Most of these thrust-bounded S-folds and Z-folds are southwesterly-overturned. In contrast with Fortune Harbour Formation strata disposed by the displaced syncline in Northwest Arm and Southeast Arm (Figure 10), the turbidite succession of the Moores Cove Formation is relatively complete in the core of the Cottrells Cove Syncline. However, tectonically adjacent anticlines are offset by D_1 imbricate thrusts (Figure 31), which are preferentially developed near the Fortune Harbour–Moores Cove transition. Near the village of Cottrells Cove, Z-shaped F_1 asymmetrical folds occur in thrust sheets containing mainly inverted stratigraphic sections; whereas, S-shaped F_1 asymmetrical folds are found in structurally repeated sections of right-way-up, northeast-dipping rocks (Figure 8).

On the northeast limb of the Cottrells Cove Syncline, the shear sense of thrust faults is the same as the shear sense of the parasitic folds; consequently, the differential displacement of bedding on the long limbs of parasitic Z-folds within thrust sheets is consistent with hanging wall-up movements on adjacent thrust faults. However, on the southwest limb of the Cottrells Cove Syncline, the shear sense of thrust faults opposes the shear sense of offset parasitic folds. Thus, the differential displacement of bedding on the long limbs of parasitic S-folds within a thrust sheet appears to indicate hanging wall-down movements rather than hanging wall-up movements. This could be explained, however, if F_1 folding preceded D_1 thrusting in this area. In any case, the S_1 slaty cleavage in northeast-dipping thrust sheets intensifies toward the bounding D_1 faults regardless of whether the contained F_1 fold is S-shaped or Z-shaped. Moreover, where juxtaposed units are biostratigraphically or isotopically dated, the hanging wall sequence of these northeast-dipping faults are known to contain strata that are older than those in the footwall sequence (Figure 8).

Significance of D_1 Thrusts with Opposing Dips

A common map pattern in central Notre Dame Bay shows northwest-trending D_1 thrust faults having opposing dips on the northeast and southwest flanks of regional F_1 folds (Figure 32). This is most simply explained by having F_1 folds of D_1 thrusts or, alternatively, D_1 thrusts of F_1 folds. In the first interpretation (Figure 32a), an antiformal pericline culminates to create a structural window through a single thrust fault to expose the footwall sequence. On one limb of the antiform, the thrust fault displays its original reverse sense of shear; whereas, on the opposing limb, the thrust is bodily rotated about the fold axis to display an apparent sense of normal fault displacement.

In the favoured second interpretation (Figure 32b), two oppositely-dipping faults display an original reverse sense of shear and both faults displace the limbs and hinge zone of an intervening conjugate anticline. The southwest-dipping reverse fault crosscuts the northeast-dipping reverse fault to form a ‘pop-up’ thrust zone which terminates to the northwest and to the southeast. Fault triangle zones are kindred structures to ‘pop-up’ thrust zones (Figure 32b).

In many places, D_1 reverse faults having the same strike but opposing dip directions confront each other (e.g., Figure 31) and, where one group of dipping faults crosscuts another group of dipping faults, they form a Y-type of structure (e.g., Figure 11). However, in some areas, the “root zone” or stem of the resulting Y-structure has a vertical dip. Overall, the V-shaped “pop-up” thrust zone at the top of the Y-structure joins with the vertical stem to produce a tectonic feature which resembles a symmetrical positive flower structure. Because many of these reverse faults have a strike-slip component of offset (with opposing sinistral and dextral shear senses), some “popped-up” thrust blocks were also laterally displaced. This means that certain “pop-up” thrust zones were ejected towards or away from the viewer of the cross section of the “pop-up” structure (Figure 32b), and this may explain some of the extreme plunge variation seen in conjugate F_1 folds within D_1 thrust sheets (Figure 32b).

D_2 Deformation

General Statement

The D_2 phase of regional deformation caused most of the widespread northeast-trending structures in central Notre Dame Bay. D_2 deformation produced regional F_2 anticlines and synclines, major oblique-slip D_2 fault zones and widely developed S_2 slaty cleavage.

Throughout the map area, northeast-trending thrust and reverse faults of D_2 age generally formed as bipolar structures in association with bivergent F_2 conjugate folds. In this regard, they mirror the D_1 structures. Major D_2 fault zones in which southeast-dipping imbricate thrust sheets confront northwest-dipping imbricate thrust sheets are also a characteristic feature of the region.

Southeast-dipping D_2 thrusts and Z-shaped F_2 folds predominate, especially in the eastern part of the area surveyed (e.g., Figure 24). In contrast, northwest-dipping D_2 thrusts and S-shaped F_2 folds increase in abundance toward the west. This reflects the position of the map area on the northwest-trending, S-folded, middle limb and the northeast-trending, Z-folded, long limb of a regional oroclinal fold, the Notre Dame Bay flexure (Figure 33).

Throughout this flexure, and particularly on the northeast-trending limbs of this Z-shaped structure, D_2 overprinting caused significant modification and regional reorientation of the northwest-trending D_1 structures (e.g., Figure 29). In places, variably doubly plunging F_2 folds deform lithotectonic sequences cut by D_1 reverse faults. In strati-

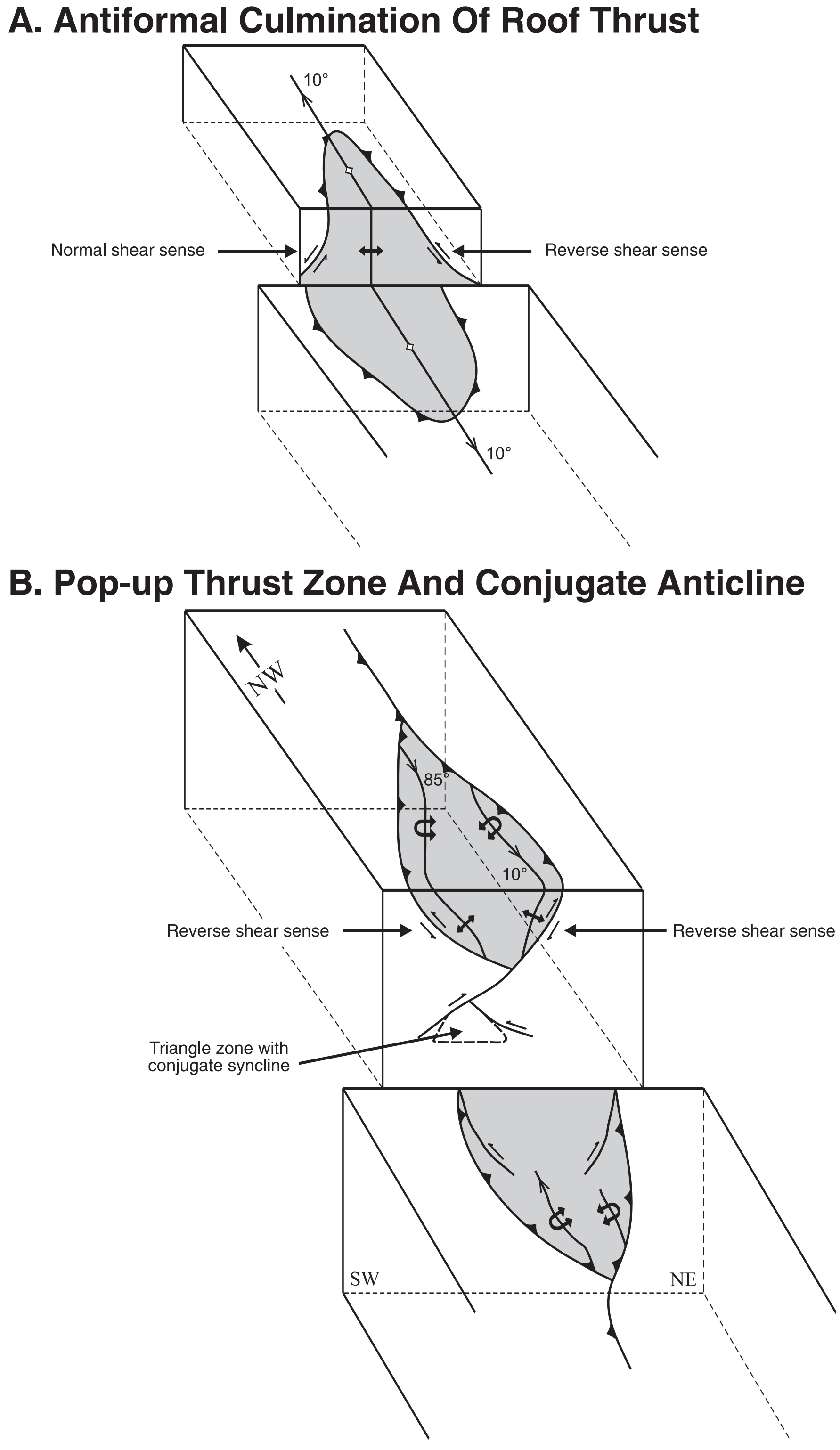


Figure 32. *Block diagrams illustrating two interpretations of northwest-trending D_1 thrust faults that dip in opposing directions either side of a northwest-trending F_1 regional fold. A) Gently doubly plunging F_1 antiformal culmination of a D_1 thrust. Note that, on one fold limb, the folded D_1 thrust fault shows an apparent normal sense of displacement (hanging wall moves downward). B) Variably plunging F_1 conjugate anticline displaced by bipolar D_1 thrust faults to create a pop-up thrust zone. Note that, because the northeast-directed fault crosscuts the southwest-directed fault, the D_1 tectonic panel containing the F_1 fold terminates in both directions along strike but it is not a thrust duplex.*

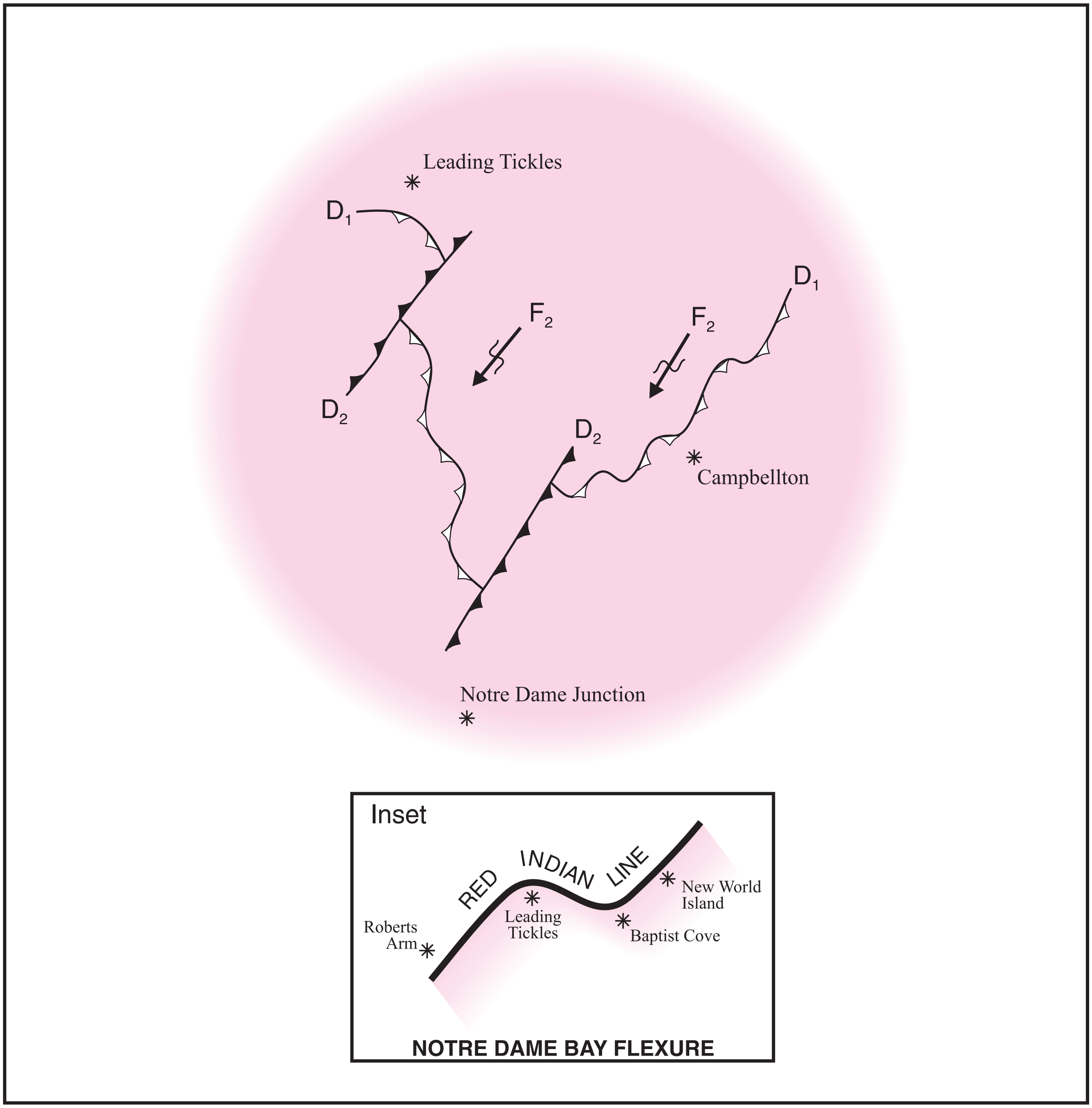


Figure 33. *Tectonic sketch map showing the eastern and middle limb of the Z-shaped Notre Dame Bay flexure and the approximate position of three geographic locations in the area surveyed. A representative D_1 thrust outlines part of this oroclinal fold. It also generally illustrates the distribution of the main types of F_2 folds and D_2 thrusts, and schematically depicts their effects on the regional structure.*

graphically continuous successions within these northeast-elongated domes and basins, vertical S_2 cleavage commonly crenulates a more gently dipping S_1 slaty cleavage. Internally, some of these complex D_2 structures preserve segments of D_1 thrust sheets and structurally detached parts of periclinal F_1 folds or they contain the tightly refolded limb of a F_1 fold nappe (e.g., Folded Thrust Duplex III in Figure 25).

Reverse faults of D_1 and D_2 age are mapped to coalesce near D_2 tracts of tectonically straightened strata or within D_2 mylonite belts adjacent to major northeast-trending fault zones. Platey rocks with composite S_1/S_2 foliation are observed to pass laterally and gradationally into embrittled tectonic melange. The block-in-matrix texture of individual D_2 melange zones is irregularly developed in small discrete patches and it is typically less than a metre thick.

Examples of D_2 Structures

Many of the major folds and fault zones which are mapped regionally throughout the interior of the area surveyed are superbly exposed as they cross the islands, fjords and coastal headlands of central Notre Dame Bay. On the eastern part of the Fortune Harbour peninsula and on islands in the western Bay of Exploits, the Notre Dame Subzone's Cottrells Cove Group passes from the northwest-striking middle limb to the northeast-striking long limb of the Notre Dame Bay flexure. On the Duck and Grassy islands, where the Fortune Harbour and Moores Cove formations trend northeastward, D_2 structures are present (Figure 9). They are generally representative of those in the eastern part of the area surveyed.

Southeast-dipping D_2 faults and northwesterly-overturned F_2 folds are well developed in this part of the Cottrells Cove Group (Figure 9). A southeast-dipping thrust sheet of the Fortune Harbour Formation, which is bounded by D_2 reverse faults, contains mainly inverted volcanic and sedimentary successions that are located on the long limbs of a large Z-shaped F_2 fold. A major F_2 anticline, which controlled the disposition of a right-way-up, southeast-dipping succession of the Moores Cove Formation in the southeasterly adjacent thrust sheet, was apparently displaced by an intervening D_2 reverse fault (Figure 9). Both thrust sheets are displaced northwestward along the D_2 Chanceport Fault and placed structurally above a southeast-facing panel of the Sweeney Island and Western Head formations of the Moretons Harbour Group. Thus, two tectonic panels of the Cottrells Cove Group and one tectonic panel of the Moretons Harbour Group comprise a southeast-dipping, non-depositional sequence that faces alternatively southeastward and northwestward.

Regional D_2 structures in the Exploits Group, the Badger Group and the Botwood Group occur south of the Red Indian Line and are characteristic of Exploits Subzone rocks in the central Notre Dame Bay region. In the South Arm of New Bay, for example, the Charles Brook member of the New Bay Formation, the Saunders Cove Formation and the Little Arm East, Pushthrough and Pleasantview members of the Tea Arm Formation are affected by northeast-trending D_2 reverse faults. These structures separate S-shaped pairs of northeast-plunging F_2 folds (Figure 13). Where the D_2 faults that bound the folded and thrusted panels dip northwestward, reverse displacements place older strata above younger strata. Where the bounding D_2 faults dip southeastward, younger-over-older translations are recorded.

Major F_2 folds in the Brooks Harbour and Saltwater Pond members of the New Bay Formation are dominantly northeast-plunging from Little Indian Cove and Bills Point in the South Arm of New Bay to High Grego Island in the Bay of Exploits (O'Brien, 1990; Figure 3). In places, the Lawrence Head Formation, the Strong Island chert and the Lawrence Harbour Formation also outline this fold train, which contains mostly open and upright structures over a strike length of approximately 13 km. Despite carrying a strong, axial planar, subvertical, S_2 slaty cleavage, such F_2 folds do not possess a well-developed sense of fold vergence. However, overturned F_2 asymmetrical folds with steeply southeast-dipping S_2 cleavage and a strong sense of vergence were developed toward the northwest and southeast ends of the fold train in the middle–upper Exploits Group. These fold structures are located near regional D_2 reverse faults that probably underlie the Southwest Arm of New Bay (Figure 13) and the Ships Run of the Bay of Exploits (Figure 9), respectively.

Though F_2 fold structures in the D_2 fold train are regionally extensive, they deform even larger sized F_1 folds and allied D_1 thrust faults. For example, in the volcanic and sedimentary rocks of the lower Exploits Group in the Tea Arm area (Figure 13), F_2 major folds and associated D_2 reverse faults deform a regional system of D_1 imbricate thrusts (the Paradise fault zone) and the core and limb of a large F_1 fold nappe (the Tea Arm anticline). A smaller example of fold interference between F_2 and F_1 structures occurs in the footwall of a southwest-directed D_1 overthrust in the Paradise fault zone (Figure 29). Regional fold interference patterns are also present in the Ritters Arm–Sunday Island area, where map-scale F_2 folds refold a large F_1 syncline in the turbidites of the middle Exploits Group (O'Brien, 1990; O'Brien *et al.*, 1997). Overprinting relationships between smaller F_2 and F_1 folds are seen near the hinge zone of this syncline (Figure 28). In the vicinity of Phillips Head Pond, the steeply northeast- plunging, S-shaped F_2 open folds of D_1 reverse faults that developed along the faulted margin of the Shoal Arm Formation also extend into the tectonically adjacent succession of the middle Exploits Group (O'Brien, 1993a). In general, the interference of northwesterly overturned F_2 folds with southwesterly-overturned F_1 folds does not produce regional scale stratigraphic repetition of the Exploits Group, as both sets of these dominant folds have the same sense of fold asymmetry within the low-strain parts of thrust sheets.

In the Little Northwest Arm of New Bay, southeast-dipping D_2 reverse faults place Arenig–Llanvirn strata from the upper Exploits Group structurally above inverted Ashgill strata belonging to the lower Badger Group (Figure 25). Here, near a splay of the Northwest Arm Fault (Figure 11), D_2 faults are also responsible for imbricating the basal and lower parts of the Point Leamington Formation of the Badger Group. In Little Northwest Arm, the southeast-dipping D_2 fault structures separate D_2 tectonic panels which contain Z-shaped F_2 folds (Figure 25). Within several panels, D_1 structures are observed to be overprinted by these F_2 folds. Also, in spectacular coastal exposures, the D_2 reverse faults displace D_1 thrust duplexes that are developed within the

Lawrence Harbour Formation. A similar type of tectonism affects Badger Group strata in the eastern part of the area surveyed. In the Ashgill succession of the Campbellton greywacke and Lewisporte conglomerate, steeply southwest-plunging F_2 folds refold southeast-plunging F_1 folds and deform southwest-dipping D_1 reverse faults in the region west of Indian Arm Brook (O'Brien, 1992b).

In the southern part of the area surveyed, the Lawrenceton and Wigwam formations of the Botwood Group, and their subunits, are disposed about regional F_2 periclinal folds (Dickson *et al*., 1995). The F_2 plunge is generally southwestward throughout most of the type area of the Botwood Group; however, a northeastward F_2 plunge is more common adjacent to the Northern Arm Fault (O'Brien, 1993a). In places, volcanic and sedimentary rocks of the Botwood Group are folded by F_2 asymmetrical folds and carry penetrative S_2 cleavage lying parallel to steeply southeast-dipping reverse faults (Figure 24). Imbricate D_2 thrusts were preferentially developed near the margins of the Botwood Group but such structures are also observed where detached slices of the Botwood Group occur within older units of the Exploits Subzone (Figure 18). Although inverted and right-way-up successions of the Wigwam Formation are present in the town of Lewisporte, the stratigraphic succession of the Lawrenceton and Wigwam formations is right-way-up in most of the larger thrust sheets of the Botwood Group (e.g., the Norris Arm North panel in Figure 22b).

In areas where the tectonic panels of Silurian terrestrial strata trend northwestward, the bounding D_1 imbricate fault zones are obviously crosscut by regional D_2 structures (O'Brien, 1993a). However, in places where the Botwood imbricate thrust stack trends northeastward, it is difficult to distinquish reoriented D_1 structures from the overprinting D_2 structures. In this regard, it is noteworthy that thrust slices of the Botwood Group trend west-northwesterly east of Browns Arm but northeasterly near Burnt Bay, Peters Pond and Norris Arm bottom (Dickson *et al*., 1995).

During D_2 deformation, the originally vesicular basalts of the Lawrenceton Formation were metamorphosed to chlorite schists. Relatively fresh metabasites are observable near Emily Cove and in the northern part of the town of Lewisporte; pyritic sericite–chlorite schist is present on the southern end of Freak Island. In this region, southeast-dipping S_2 schistosity transposes stratigraphic layering (and S_1 cleavage in places) and is crosscut by altered quartz-feldspar porphyry sheets from the Loon Bay batholith. Moderately pitching L_2 extension lineations are locally seen in microconglomerate beds in the Wigwam Formation on Rice Island [E643550 N5458300].

F_2 Periclinal Folds

In the central Notre Dame Bay, it is common to have regional trains of northeast-trending F_2 folds cross large-strain D_1 structures at the northwest-trending margins of competent units, and to observe these F_2 folds increase their amplitude and wavelength within negligibly strained strata in adjacent incompetent units. An example of this phenomenon occurs in the New Bay Pond area, where the well-bedded sedimentary rocks of the Caradoc Shoal Arm and Ashgill Gull Island formations comprise the lower thrust plate immediately below the southwest-dipping D_1 mylonites of the West Arm Brook Thrust and the presumed early Ordovician basalts of the Wild Bight Group in the upper thrust plate (Swinden and Jenner, 1992; Dickson *et al*., 1995; Figure 3). In the massive volcanic rocks of the D_1 hanging wall sequence, F_2 folds are absent or very weakly developed. The S_1 mylonites derived from these basalts illustrate spaced S_2 crenulations and associated F_2 microfolds at the West Arm Brook Thrust. However, in Upper Ordovician strata within the northwest-striking D_1 footwall sequence, vertical to steeply southwest-plunging F_2 folds display axial planar S_2 slaty cleavage for several kilometres northeast of the openly F_2-folded West Arm Brook Thrust.

Near the margin of the Point Leamington basin, in the area between Martin Lake and Askel Lake, a similar spatial relationship exists between the presence and shape of F_2 folds and their position in a D_1 strain gradient. There, a northwest-trending D_1 thrust places massive plutonic rocks from the Cambro-Ordovician South Lake Igneous Complex above a footwall sequence that contains a continuous Upper Ordovician succession from the lower sandstone turbidites of the Point Leamington Formation to the stratigraphically lowest lenticle of Goldson conglomerate (O'Brien, 1992a; MacLachlan, 1998). Northeast-trending F_2 structures openly fold the bounding D_1 thrust and a bedding-parallel S_1 cleavage in tectonicaly adjacent strata of the Point Leamington Formation. Farther northeast along the F_2 axial traces, the F_2 cross folds become tighter, dominate over the F_1 structures, and dispose higher parts of the Ashgill stratigraphic succession. Periclinal F_2 folds display less overprinted S_1 cleavage and carry a more pervasive S_2 cleavage, as they propagate across the southwest margin of the Point Leamington basin.

Between Stanhope Pond and Southwest Pond, near the D_1 reverse fault at the margin of the Botwood Group (Figure 18), a strong southwest-dipping S_1 cleavage in the Lawrenceton Formation is regionally folded and locally crenulated by F_2 folds. Higher in the D_1 hanging wall plate, the S_1 foliation gradually dies out in a right-way-up succession of competent mafic and felsic volcanic rocks. To the southwest, down the direction of F_2 plunge in the overlying sedimentary rocks of the lower Wigwam Formation, major folds of D_2 age increase in prominence and sporadically display northeast-trending S_2 slaty cleavage. Farther southwest, in the upper Wigwam Formation, the axial traces of F_2 periclinal folds themselves outline large S-shaped fold structures (O'Brien, 1993a), possibly implying late stage movements along adjacent fault zones in the Botwood Group.

In the area between St. John's Bay and Indian Arm, the Exploits Group and the Dunnage Melange pass from the northwest-striking limb of the Notre Dame Bay flexure onto one of its northeast-striking limbs (Figure 34). In this

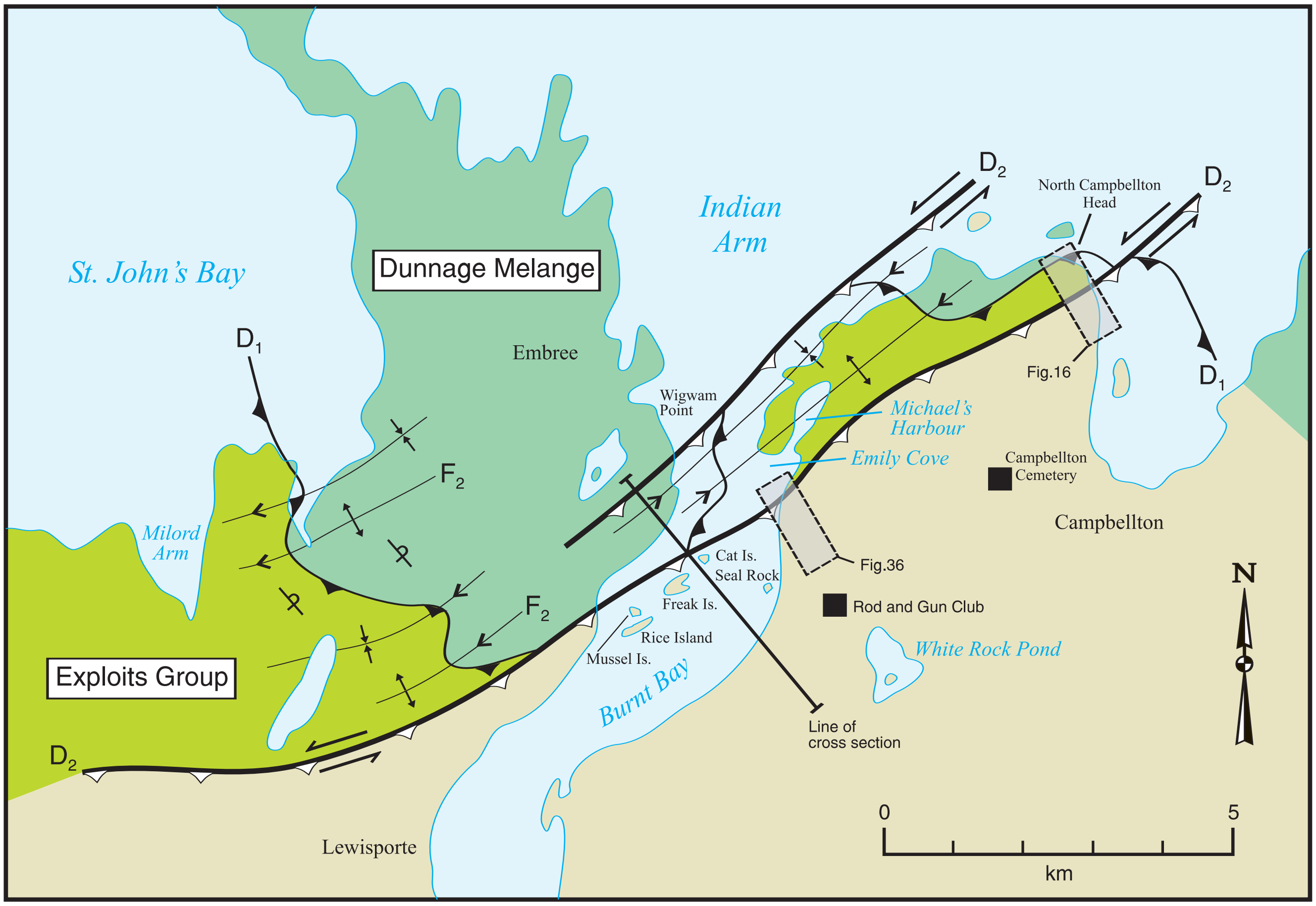


Figure 34. *Simplified structural map of the St. John's Bay–Indian Arm area, where a regionally folded D_1 thrust placed the Exploits Group structurally above the Dunnage Melange. Two sinistral oblique-slip D_2 reverse faults dip southeastward, displace the regional D_1 thrust, and are developed on the limbs of northeast-trending, gently doubly-plunging F_2 periclinal folds. Locations of Figures 16 and 36 are illustrated. Also denoted is the line of cross section for Figure 35.*

region, a D_1 thrust, which originally placed the Exploits Group structurally above the Dunnage Melange, was deformed during D_2 deformation to create large, northeast-trending, periclinal F_2 folds. Although most F_2 fold pairs outline Z-shaped structures, an important S-shaped F_2 fold pair is present near Michaels Harbour (Figure 34). A periclinal plunge depression between Campbellton and Burnt Bay preserves the upper D_1 thrust plate carrying the Exploits Group in a D_2 structural basin. In contrast, a periclinal plunge culmination between Milord Arm and Burnt Bay creates a D_2 structural dome and hence a window into the lower D_1 thrust plate carrying the Dunnage Melange. Southeast-dipping D_2 reverse faults are situated on the limbs of the gently plunging F_2 periclines (Figure 16 inset and Figure 34).

F_2 Asymmetrical Folds

In eastern Notre Dame Bay, gently northeast- and southwest-plunging F_2 folds with a well developed sense of fold asymmetry are most commonly overturned to the northwest (e.g., van der Pluijm, 1986). In the North Campbellton area, major southwest-plunging F_2 fold pairs and an intervening southeast-dipping D_2 oblique-slip reverse fault point to this same general direction of D_2 tectonic transport (Figure 16). From northeast to southwest along its trace, this D_2 reverse fault is mapped to ramp upwards from the Exploits Group, cut through the overlying Luscombe shale and pass into the Campbellton greywacke of the Badger Group. The D_2 reverse fault lies several hundred metres southeast of an asymmetrically folded D_1 thrust that had placed the Exploits Group above the Dunnage Melange. It is along this older structure that most of the high-strain regional deformation is localized. On North Campbellton Head, tectonically-straightened strata on the long limbs of asymmetrical folds have strong extension lineations which pitch moderately to steeply on composite S_2/S_1 cleavage. The lineations indicate an oblique dip-slip movement on these southeast-dipping ductile faults.

Minor structures of D_1 and D_2 age occur in strata of the Exploits Group and the Dunnage Melange on the northwest side of the D_2 reverse fault and are found in strata of the Badger and Exploits groups on the southeast side of the fault. South of North Campbellton Head, near the D_2 fault structure, S_1 cleavage surfaces and F_1 minor folds are reoriented from a northwesterly strike in the hanging wall and footwall sequences into a northeasterly trend adjacent to the D_2 fault structure (Figure 16). In contrast, S_2 cleavage dips consistently southeastward in this area. In the D_2 footwall sequence, the New Bay Formation of the middle Exploits Group and the Strong Island chert of the upper Exploits Group are present in the asymmetrically folded D_1 thrust slices above the Dunnage Melange, although the Lawrence Head Formation is completely omitted. The hanging wall sequence of the D_2 reverse fault appears to contain an unfaulted stratigraphic succession from the uppermost Exploits Group to the lowermost Badger Group. Based on bedding–S_1 cleavage intersections, these rocks, along with a swarm of Badger Group-hosted gabbros, occur in the hinge zone of a northeasterly-overturned F_1 anticline (Figure 16), whose northwest-striking limb is presumably located offshore under Campbellton Harbour.

In certain parts of the map area, superimposed F_2 asymmetric folds have a highly variable geometry which reflects the orientation and shape of pre-existing F_1 fold structures. On Big Island in the Western Arm of New Bay, a steeply-plunging F_2 antiform–synform pair crosses both limbs of an inclined F_1 syncline cored by the Randels Cove conglomerate (O'Brien, 1990). Thus, there are positions along the said F_2 fold axial traces in Western Arm where the F_2 antiform is a syncline and where the paired F_2 synform is an anticline. Similarly, within the exposures of the Charles Brook member of the New Bay Formation on Thwart Island, the vertical axial surface of the regional F_1 Tea Arm Anticline outlines a northeast-trending, vertically plunging F_2 fold (Hughes and O'Brien, 1994). However, in places, the shape of this northeastward-closing F_2 vertical fold defines neither an upward-closing antiform nor a downward-closing synform. On the west coast of Thwart Island, immediately north of the F_1 axial trace of the Tea Arm Anticline, the superimposed F_2 fold is a neutral anticline. North of Seal Rocks, for a short distance south of the folded vertical beds in the F_1 hinge zone, the same F_2 fold closes sideways as a neutral syncline.

A strongly preferred direction of F_2 overfolding and D_2 overthrusting is seen in many of the regional fault zones in central Notre Dame Bay (e.g., northwestward near the D_2 Chanceport Fault in the Bay of Exploits; Figure 9). However, southeast-dipping D_1 thrusts (having strongly asymmetric F_1 folds in the hanging wall or footwall plates) are also found near these northeast-trending, southeast-dipping D_2 fault zones (Figure 33). They are now observed to be overturned in the same direction as the northwesterly-overturned F_2 folds (Figure 16) or, alternatively, in an opposing direction (Figure 25). In the Swan Islands, in the footwall sequence of a D_2 overthrust in the Lukes Arm fault zone, the phenomenon of pervasive northwest-directed D_2 structures overprinting more restricted southeast-directed D_1 structures is well displayed by the turbidites of the Moores Cove Formation (Lafrance, 1989). On Tinker Island, pillowed basalts are imbricated with strongly deformed turbidites and other volcanic rocks in southeast-dipping D_2 thrust sheets and, near Swan Island Harbour, they are also seen to be mylonitized adjacent to D_2 oblique-slip reverse faults. Both the relict southeast-directed D_1 structures and the preferred northwest-directed D_2 structures were reworked within the northeast-trending belts of tectonic melange found discontinuously throughout the Hornet Island area (Lafrance, 1989; O'Brien, 1991b).

D_2 Reverse Faults

The area east of Lewisporte represents an example of a northeast-trending belt of D_2 reverse faults and imbricate D_2 thrust sheets superimposed on a similarly oriented system of D_1 thrusts (Figure 35). Lithological units of the Botwood Group, the Badger Group, the Exploits Group and the Dunnage Melange each occur in several large tectonic panels that cross Burnt Bay and extend into the Campbellton area. Generally, within individual thrust sheets, right-way-up strata dipping southeastward are more commonly observed than the inverted successions of northwest-younging strata. Consequently, a down-plunge view of this belt contains more S-shaped F_2 folds than it does Z-shaped F_2 folds, a unique feature given its position on the eastern long limb of the Notre Dame Bay flexure.

Southeast-dipping D_2 reverse faults and northwesterly-overturned F_2 folds are depicted in a generalized cross section, viewed looking northeastward, from Wigwam Point on Burnt Bay to White Rock Pond, some 2.5 km farther southeast (Figures 34 and 35). Both the Z-shaped F_2 folds and the S-shaped F_2 folds are displaced by southeast-dipping D_2 reverse faults. In places where a Z-folded panel is juxtaposed against a S-folded panel, strata in the D_2 hanging-wall sequence become younger in the opposite direction to strata in the adjacent D_2 footwall sequence (Figure 35). However, in the Wigwam Point–White Rock Pond area, it is more common to find a sequence of S-folded panels thrust faulted against one another. Therefore, most of this D_2 lithotectonic sequence is southeast-facing, albeit stratigraphically dismembered.

In the Burnt Bay area, the displacement sense of the D_2 reverse faults opposes the shear sense of the S-shaped F_2 folds between such faults. The geometrical relationship is similar to that described for D_1 thrusts and F_1 folds located on the southwest limb of the Cottrells Cove Syncline (Figure 8).

The fold-and-thrust belt underlying Burnt Bay has D_1 structures which show the same general direction of tectonic transport as the above mentioned D_2 structures. Superposition of a variety of D_2 tectonic structures on thrust-related D_1 minor structures can be observed in several locations. Northeast of Lewisporte, along the coast line south of Emily Cove (Figure 36), examples of overprinted D_1 structures

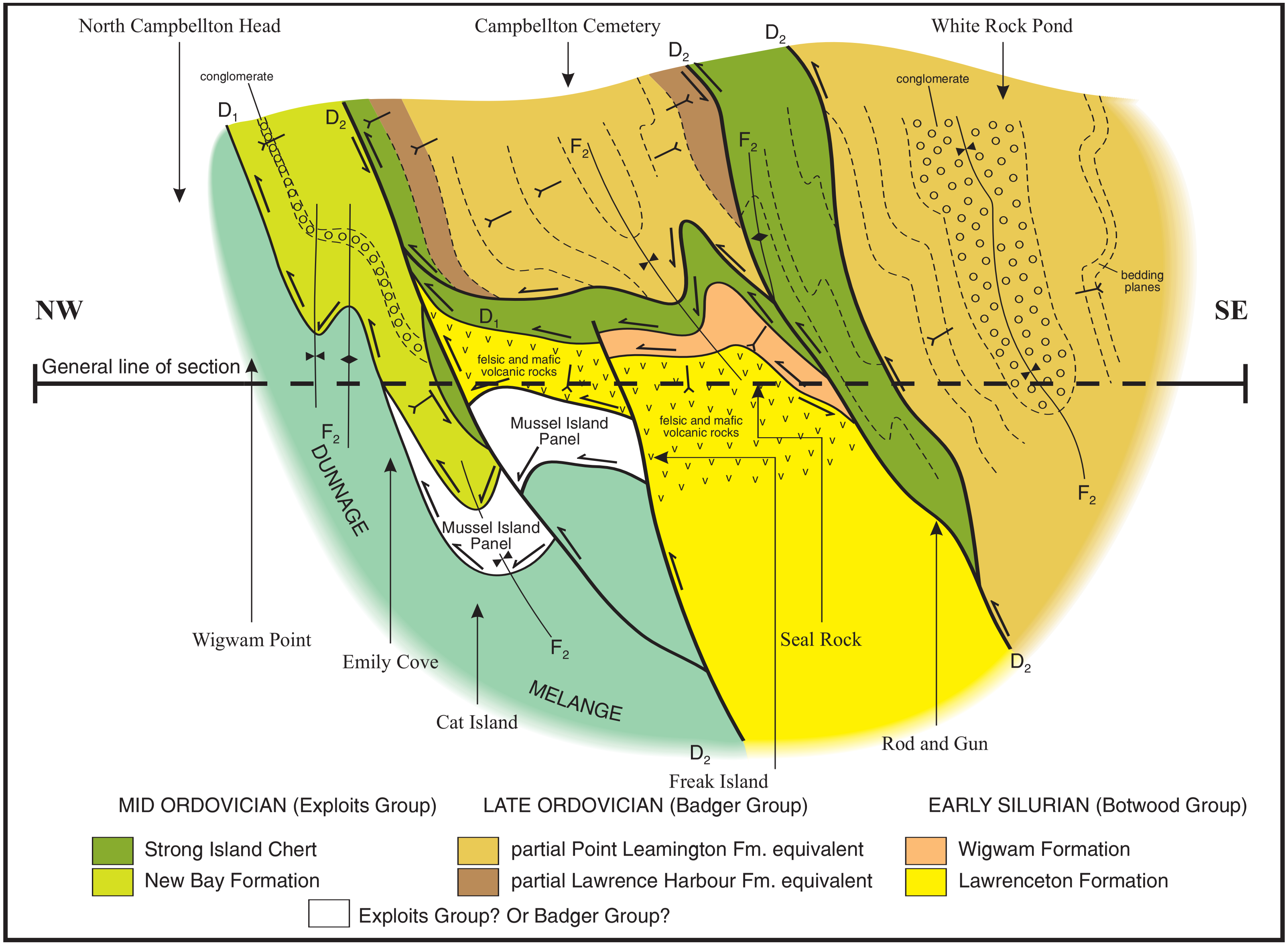


Figure 35. *Generalized northwest–southeast cross section across Burnt Bay, from Wigwam Point to White Rock Pond, depicting the inferred geometry of regional tectonic panels of the Dunnage Melange, the Exploits Group, the Badger Group and the Botwood Group. Lithostratigraphic and structural data from the Campbellton area projected several kilometres southwestward into the Burnt Bay section. Note the D_2 reverse offset of D_1 thrust sheets and the open F_2 folding of D_1 thrust faults. Vertical scale exaggerated; position of line of section from Figure 34 is approximate. Section displays rocks above and below the present erosional surface without distinction.*

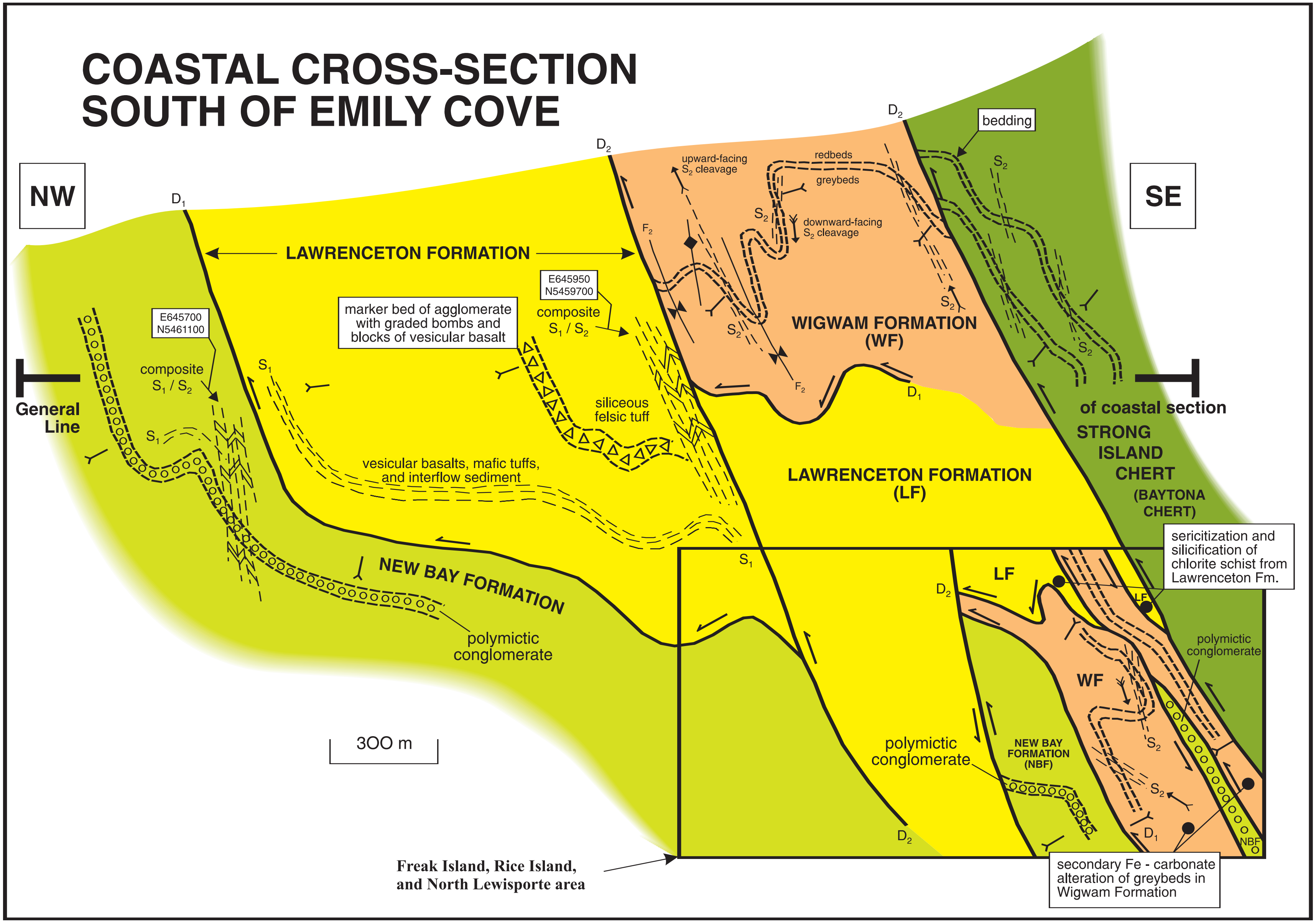


Figure 36. *Detailed cross section of the Botwood Group and tectonically adjacent map units south of Emily Cove on Burnt Bay. Lithological units of the Lawrenceton and Wigwam formations outline S-shaped F_2 folds within an imbricate D_2 thrust stack. Also depicted is an interpretation of the relationships of observed D_1 and D_2 minor structures to major D_1 thrust and D_2 reverse faults within the lithotectonic sequence. Structural data from Rice and Freak islands and North Lewisporte is projected into the plane of the Emily Cove cross section. Vertical scale is non-uniformly exaggerated. See Figure 34 for section location.*

include those seen along the Botwood Group–Exploits Group boundary at [E645700 N5461100] and along the Lawrenceton Formation–Wigwam Formation boundary at [E645950 N5459700].

Where lithological units generally strike northeastward, strata-parallel D_1 thrust faults, F_1 minor folds and S_1 cleavage surfaces are still recognizable over large tracts of ground in the Lewisporte region (O'Brien, 1992b). They can be utilized to map out the coaxial F_2 folds which overprint these D_1 structures (e.g., Figures 16 and 36). Within the particular thrust sheets underlying the Burnt Bay–Indian Arm area, the presence of F_2 folds is independently confirmed by conspicuous conglomerate horizons in middle and upper Ordovician units, folded pretectonic gabbroic intrusions in middle and upper Ordovician units, and regional variations in the stratigraphical facing directions of these and other units (e.g., Figures 16, 35 and 36).

Major S-shaped F_2 folds of the relict D_1 thrust stack underlying the area east of Lewisporte vary from open to tight, when viewed looking northeastward down the F_2 fold plunge (Figure 35). However, in a few localities, gently plunging, Z-shaped F_2 folds are defined by D_1 thrust faults (*also see* Figure 34). Some of them plunge southwestward. Where weakly developed, F_2 folds are restricted to the vicinity of the deformed D_1 boundary thrust (Figure 35). Yet, in many other areas, tight F_2 folds with strong S_2 cleavage affect large portions of the original D_1 thrust sheets.

D_2 reverse offset of D_1 thrust sheets means that certain parts of these tectonic panels of Ordovician and Silurian rocks are bounded by D_2 structures; whereas, adjacent parts of the same panels are affected by D_1 structures. Moreover in some locations, especially near the margins of the Silurian Botwood Group, D_1 thrusts may be locally coplanar with the southeast-dipping D_2 reverse faults. These phenomena may partly account for the occurrence of both older-over-younger and younger-over-older tectonic movements within the southeast-dipping, composite D_1/D_2 thrust stack.

Relationships Between Northeast-trending D_1 and D_2 Structures

Along the Burnt Bay coast and northeastward toward Campbellton, the Botwood Group is structurally underlain by the New Bay Formation of the Exploits Group on its northwestern margin and is structurally overlain by the Strong Island chert of the Exploits Group on its southeastern margin (Figure 36). Internally, a mainly right-way-up sequence of the Wigwam Formation is in direct D_1 or D_2 faulted contact with a mainly right-way-up sequence of the underlying Lawrenceton Formation. Nevertheless, structural slices of each Botwood Group formation contain relatively thin lithostratigraphic successions which are fully representative of the lithodemes in the type area of these units (e.g., the basalt–rhyolite transition in the Lawrenceton Formation or the grey bed–red bed transition in the Wigwam Formation; Figure 36).

Thrust sheets of Lawrenceton and Wigwam strata are detached from the main belt of the Botwood Group in the northern part of the town of Lewisporte (Figure 18). There, a highly sheared and altered, southeast-dipping sliver of Lawrenceton basalt structurally overlies a right-way-up, southeast-dipping panel of the younger Wigwam Formation (Figure 36). In contrast, on adjacent islands, similar Lawrenceton basalts structurally underlie right-way-up, southeast-dipping Wigwam sandstones. Farther northwest on the mainland toward Wigwam Point, the same grey cross-bedded sandstones overthrust a panel of mainly inverted strata from the same subunit of the Wigwam Formation. In places along the strike of this southeast-dipping overthrust, these right-way-up Wigwam beds are separated from the upside-down part of the Wigwam Formation by a small tectonic wedge of the New Bay Formation. The small imbricate slices of the Exploits Group that are found within the Botwood Group lithotectonic sequence were probably detached from the structurally lowest thrust sheet of the D_1/D_2 stack underlying Burnt Bay (Figure 36).

Near the east coast of Burnt Bay, and in the Emily Cove section in particular, regionally southeast-facing, openly folded, southeast-dipping strata are typical of the D_2 thrust sheets carrying the Lawrenceton and Wigwam formations of the Botwood Group (Figure 36). In certain places, however, grey beds of the lower Wigwam Formation were more tightly folded by strongly asymmetric Z-shaped F_2 folds. Near the west coast of Burnt Bay, where southeast-dipping tectonic panels of inverted variably altered Wigwam strata are directly tectonically juxtaposed with southeast-dipping tectonic panels of right-way-up Wigwam strata, S_2 cleavage–bedding intersections indicate complex relationships between overturned folds and thrust faults in the Botwood Group.

In the town of Lewisporte, northwest-younging, upside-down Wigwam sandstones have southeast-dipping S_2 cleavage lying gentler than bedding on the long limbs of northwesterly-overturned F_2 folds (lowest WF slice in Figure 36). Here, S_2 cleavage faces structurally upward as it passes through these inverted beds; whereas in most other parts of the Botwood Group, the upward-facing S_2 cleavage dips steeper than bedding. However, in parts of the Emily Cove coastal section and especially on Rice Island, the intersection of subvertical S_2 cleavage with less steeply dipping beds (Figure 36) is not compatible with the observed northwest-younging direction of the local Wigwam succession (i.e., S_2 cleavage at this locality faces downward on bedding). Thus, minor folds in an inverted northwest-facing panel of the Wigwam Formation locally display well-developed S_2 cleavage that faces, in the structural sense, both upwards and downwards. This could be interpreted to mean that S_2 cleavage has a non-axial planar relationship with F_2 folds in certain areas underlain by the Botwood Group or it could imply that S_2 cleavage is superimposed on a F_1 fold in the Wigwam Formation.

South of Emily Cove, F_2 folds in the Lawrenceton Formation of the Botwood Group and in the adjacent New Bay Formation of the Exploits Group deform strong S_1 cleavage lying parallel to stratigraphic layering. Steeply southeast-dipping S_2 cleavage crenulates the S_1 foliation in F_2 hinge zones and gently dipping fold limbs; whereas, a composite S_2/S_1 crenulation cleavage is locally well developed adjacent to D_2 reverse faults at the boundaries of the D_2 tectonic panels (Figure 36). Downward-facing S_2 foliation in the Emily Cove section of the Wigwam Formation may indicate that F_2 folding began prior to, but then kept pace with, the development of the regional S_2 foliation and the D_2 reverse faults. It is also possible that such early D_2 deformation resulted from incipient F_2 buckling of Wigwam strata above the D_1 thrust fault which separates it from the structurally underlying Lawrenceton Formation.

Southeast-dipping D_2 and Northwest-dipping D_2 Fault Structures

Upper Black Island is representative of areas where southeast- and northwest-dipping D_2 structures overprint regional northwest-trending D_1 structures and locally reorient the older structures into the D_2 trend. The northeastern part of the island is dominated by a northwest-trending belt of southwest-dipping D_1 thrust faults and northeasterly-overturned F_1 folds (Figure 37). These affect the Llanvirn–Llandeilo and older Dunnage Melange, several Arenig–Llanvirn sedimentary and volcanic units of the Exploits Group, and the Ashgill–Llandovery turbidites of the Upper Black greywacke of the Badger Group. Deformed gabbros, which are present in all rock units, have been omitted from the structural map of Upper Black Island for the sake of clarity.

In the northwestern part of Upper Black Island, southwest-plunging, S-shaped F_2 folds and attendant northeast-trending S_2 cleavage overprint inverted sequences of northeast-younging strata within southwest-dipping D_1 imbricate thrust sheets (Figure 37). In contrast, southwest-plunging Z-shaped F_2 folds overprint similar D_1 structures in the southeastern part of the island. In central and southwestern Upper Black Island, the southwest- closing F_2 major antiform predicted by the S- and Z-shaped F_2 folds is replaced by a complex northeast-trending zone of D_2 reverse faults (see development of Phase III-type structures in Figure 26 for analogy).

The D_2 fault zone on southwestern Upper Black Island is depicted in a schematic block diagram (Figure 37), which shows the confrontation between southeast- and northwest-dipping D_2 fault structures. On its northwest margin, the Saltwater Pond member of the New Bay Formation of the Exploits Group is interpreted to have overplated the Badger Group along a dextral oblique-slip D_2 reverse fault that dips northwestward. Here, the Lawrence Head basalt, the Hummock Island limestone and the Lawrence Harbour chert and shale were completely tectonically excised, although they are all present in the D_1 thrust stack in the northeastern part of the island.

On the southeast margin of the D_2 fault zone, the Dunnage Melange is interpreted to have overplated the same tectonic panel of the Badger Group along a sinistral oblique-slip D_2 reverse fault that dips southeastward. Well-displayed D_1 structures in the unbroken coticule-rich tracts of the Dunnage Melange outline F_2 periclinal folds which plunge northeastward and southwestward (Figure 37). This D_2 structural dome in the hanging wall plate formed near the western termination of the regional D_1 thrust sheet carrying the Dunnage Melange (Figure 37 inset), and it is analogous with major D_2 structures developed between North Lewisporte and Campbellton (compare with Figure 34).

Immediately adjacent to its southeast-dipping D_2 reverse-faulted boundary with the underlying Badger Group, D_1 thrusts and F_1 folds in the Dunnage Melange (and the tectonically adjacent slice of the Exploits Group) were reoriented and the unit locally displays a composite S_2/S_1 foliation (Figure 37). In addition, the rotation of northeast-trending S_2 slaty cleavage from a vertical orientation within the F_2 periclinal dome to a gentle southeastward inclination at the D_2 reverse fault is observable in the coastal section along a small cove at the southwest end of Upper Black Island. This half-fan of cleavage occurs within the lowest part of the thrust sheet carrying the Dunnage Melange. An outcrop scale example of a similar S_2 cleavage fan has been previously reported from well-bedded chert of the Exploit Group on the southwest end of Strong Island (e.g., Plate 7 in O'Brien, 1993b). There, the vertical S_2 footwall foliation is observed to be rotated into a gentle southeastward dip beneath the hanging wall ramp of a bedding-parallel D_2 fault.

On Upper Black Island, it seems that some of the relatively incompetent units within the southwest-over-northeast D_1 thrust belt were squeezed upwards into a sheared-out antiform during regional D_2 deformation. Such rocks are represented by the Ashgill–Llandovery turbidites from the Badger Group and, possibly, the Arenig turbidites from the middle Exploits Group. The D_2 reverse faults which developed in the southwest portion of the island displaced parts of this F_2 antiform during the formation of a D_2 structural triangle zone. However, the footwall sequence beneath the overriding D_2 thrust plates was also laterally ejected from the D_2 fault zone and tectonically extruded toward the southwest. This was most likely a consequence of the different senses of oblique-slip displacement on the confronting southeast- and northwest-dipping D_2 faults (*see* Figure 37 block diagram).

Superimposed D_2/D_1 Imbricate Fault Zones with Dual Thrust Polarity

In the Leading Tickles West area (O'Brien, 1991), map units of Caradoc and Ashgill age define a northeast-trending

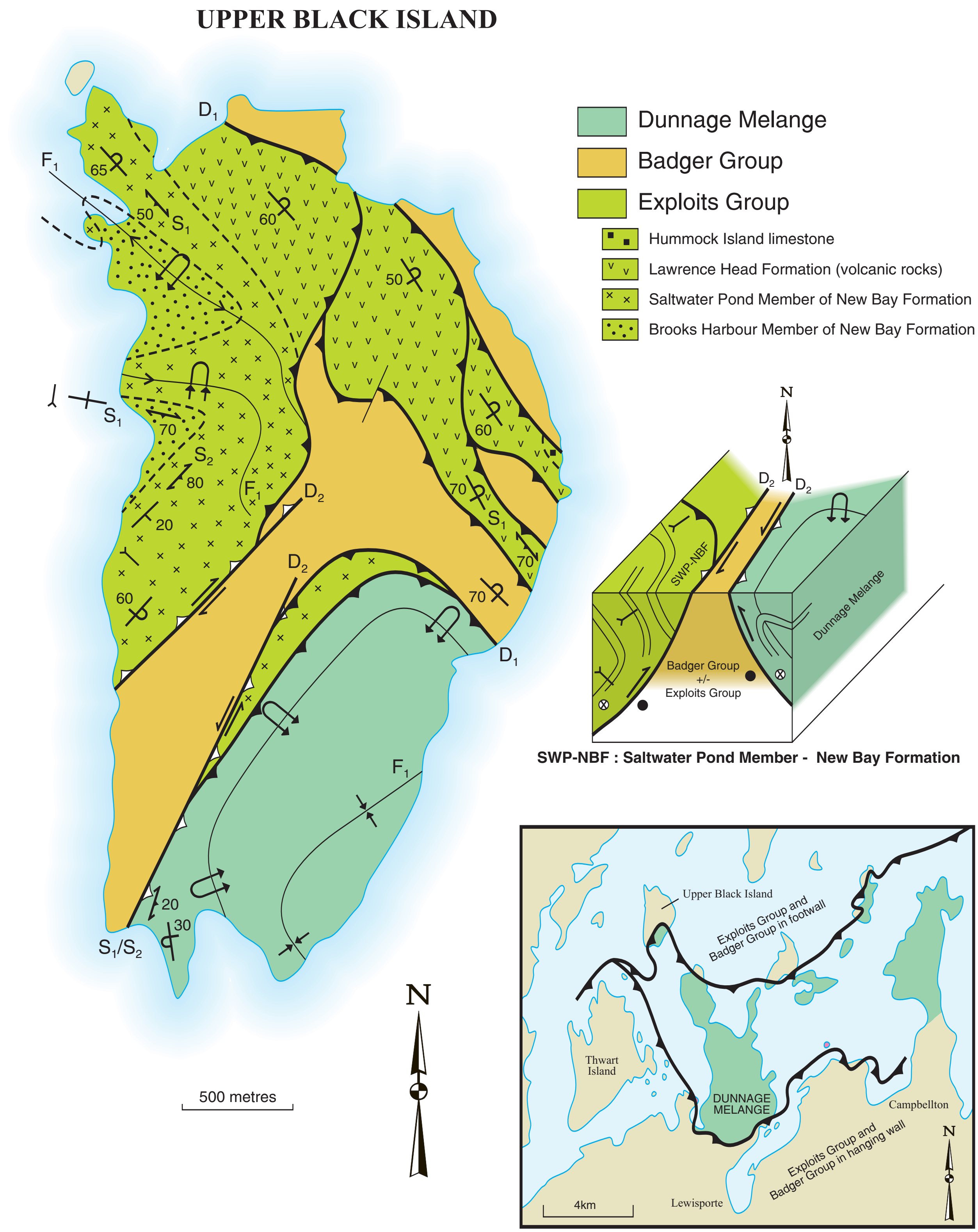


Figure 37. *Structural map of imbricate D_1 thrust sheets (and large F_1 folds) in the Badger Group, Exploits Group and Dunnage Melange on Upper Black Island. Note the differing F_2 asymmetry of superimposed F_2 folds outlined by the D_1 thrusts at group boundaries. Block diagram highlights the confrontation of southeast- and northwest-dipping D_2 faults in southwestern Upper Black Island, and the southwestward strike-slip displacement of Badger Group strata between these D_2 reverse faults. Inset depicts the regional D_1 thrust sheet containing the Dunnage Melange and the nature of its southwestern termination.*

fold-and-thrust belt located immediately southeast of the Red Indian Line. This belt displays folded D_1 fault structures that are locally crosscut by D_2 reverse faults. In this regard, the Leading Tickles West fold-and-thrust belt is geometrically similar to the belt of folded thrust duplexes found near the splay of the Northwest Arm fault zone in the outer reaches of Osmonton Arm (Figures 11 and 25).

Between Burnt Island and East Bear Cove, and also between Tinker and north Alcock islands (O'Brien, 1991), steeply southeast-dipping D_2 imbricate thrust sheets contain detached F_2 asymmetrical folds and illustrate strong S_2 cleavage. Northwest- and southeast-dipping D_1 faults are also present. On several well-exposed islands, thin highly-sheared slices of Caradoc black shale, which are bounded on both sides by southeast-dipping D_1 reverse faults, map out a tight F_2 fold train of Z-shaped folds within Ashgill turbidites. However, on Thomas Rowsell and Cull islands, the same belt of Caradoc and Ashgill strata contain northwest-dipping D_1 thrusts and southeasterly-overturned F_1 folds that are also deformed by the above-mentioned Z-shaped F_2 folds. The same can be said of Llanvirn–Llandeilo rocks in an imbricate D_1 thrust stack within the Sops Head Complex on Woody and Green islands. These mainly northwest-dipping D_1 tectonic panels are apparently truncated by a southeast-dipping D_1 thrust stack (Figure 25, *inset*). On the mainland and adjacent islands, the southeast-dipping D_1 thrust sheets contain the lowest exposed parts of the Caradoc black shale–chert succession and the upper parts of the structurally overlying Wild Bight Group; however, they are deformed by large S-shaped F_2 folds in this region.

On Red Island off Alcock Island, northwesterly-overturned F_2 folds with southeast-dipping S_2 axial planar cleavage overprint a southeasterly-overturned F_1 antiformal syncline and a small northwest-dipping D_1 limb thrust in the upper Wild Bight Group. Along strike near Jobs Cove on southern Alcock Island, a narrow tectonic sliver of probable Ashgill turbidites lying within Wild Bight strata may mark the trace of a major southeast-dipping D_2 reverse fault underlying the East Tickle. Based on comparisons with the Little Northwest Arm and North Campbellton areas, a similar D_2 structure located farther northwest probably separates the Z-shaped F_2 folded part of the D_1 thrust zone from the S-shaped F_2 folded portion of the D_1 thrust zone.

Multiple Generations of S_2 Cleavage and Antivergent F_2 Folds

Late-stage bipolar D_2 structures are discernible in the lower part of the Point Leamington Formation of the Badger Group. Near the southeast shore of Point Leamington Harbour, where the D_1 New Bay fault zone (Figure 11) is reoriented to trend northeastward, numerous southeast-dipping D_2 reverse faults are present. At this location, the New Bay fault zone may also include southeast-dipping D_1 thrusts which were folded from their original northeast-dipping attitude during D_2 deformation.

Northwest-younging Ashgill turbidites are generally right-way-up as close as 100 m to the superposed D_2/D_1 fault zone. Nearer the overthrust zone (and the upper thrust plate of Llanvirn–Llandeilo Exploits Group rocks), the Ashgill footwall sequence of the Point Leamington Formation preserves excellent F_1 minor folds and is upside-down in many places (Blewett, 1991). These F_1 folds were refolded by widespread, northwesterly-overturned, Z-shaped F_2 folds which possess a southeast-dipping, axial planar S_2 cleavage.

The stratigraphical succession of fossil-bearing Ashgill rocks was partly inverted when late-stage S_2 slaty cleavage was imposed on the northeast-striking footwall sequence of the New Bay fault zone (e.g., Plate 6 in O'Brien, 1993b). Such northeast-trending, vertical to steeply northwest-dipping S_2 cleavage lies clockwise of the southeast-dipping S_2 cleavage and crosscuts the northwesterly-overturned, Z-shaped F_2 folds. In a few localities, this late-stage S_2 cleavage is observed to be axial planar to S-shaped F_2 minor folds (Figure 38).

Thus, the younger set of F_2 minor folds and the late-stage S_2 cleavage are downward-facing structural features which locally overprint the predominant, upward-facing D_2 and D_1 structures in the footwall sequence of the D_2 New Bay overthrust. In Point Leamington Harbour, the downward-facing D_2 structures display a southeastward polarity which opposes or is antivergent to the general northwestward direction of tectonic transport associated with the earlier-formed D_2 thrusts (Figure 38). A similar late D_2 development of S_2 slaty cleavage may occur in the Exploits Group near southeast-dipping granite porphyry sheets located between Randells Cove and Diver Pond in the South Arm of New Bay within the footwall sequence of the Paradise fault zone (O'Brien, 1990). Such late D_2 features may herald the beginning of the D_3 phase of regional deformation, as they have a similar geometry to a certain group of D_3 minor structures.

D_3 Deformation

Northeast-trending D_3 minor structures are coaxially superimposed on D_2 structures; however, they are generally restricted to the major D_2 fault zones (O'Brien, 1993b). The fold vergence of F_3 minor folds and the cleavage vergence of S_3 crenulation cleavage are commonly, although not always, opposite to those of the overprinted F_2 folds and S_2 cleavage. Though sensitive indicators of late-stage thrust movements and thrust sheet readjustments, D_3 structures have not been observed to rework earlier tectonites in zones of large D_3 strain.

In the area surveyed, spaced S_3 crenulations and open F_3 folds developed with two distinctive dips or axial surface inclinations, one gentle and the other steep (compare Figures 39a and b). For example, in Late Ordovician strata of the Lawrence Harbour and Point Leamington formations, gently southeast-dipping S_3 crenulation cleavage crosscuts

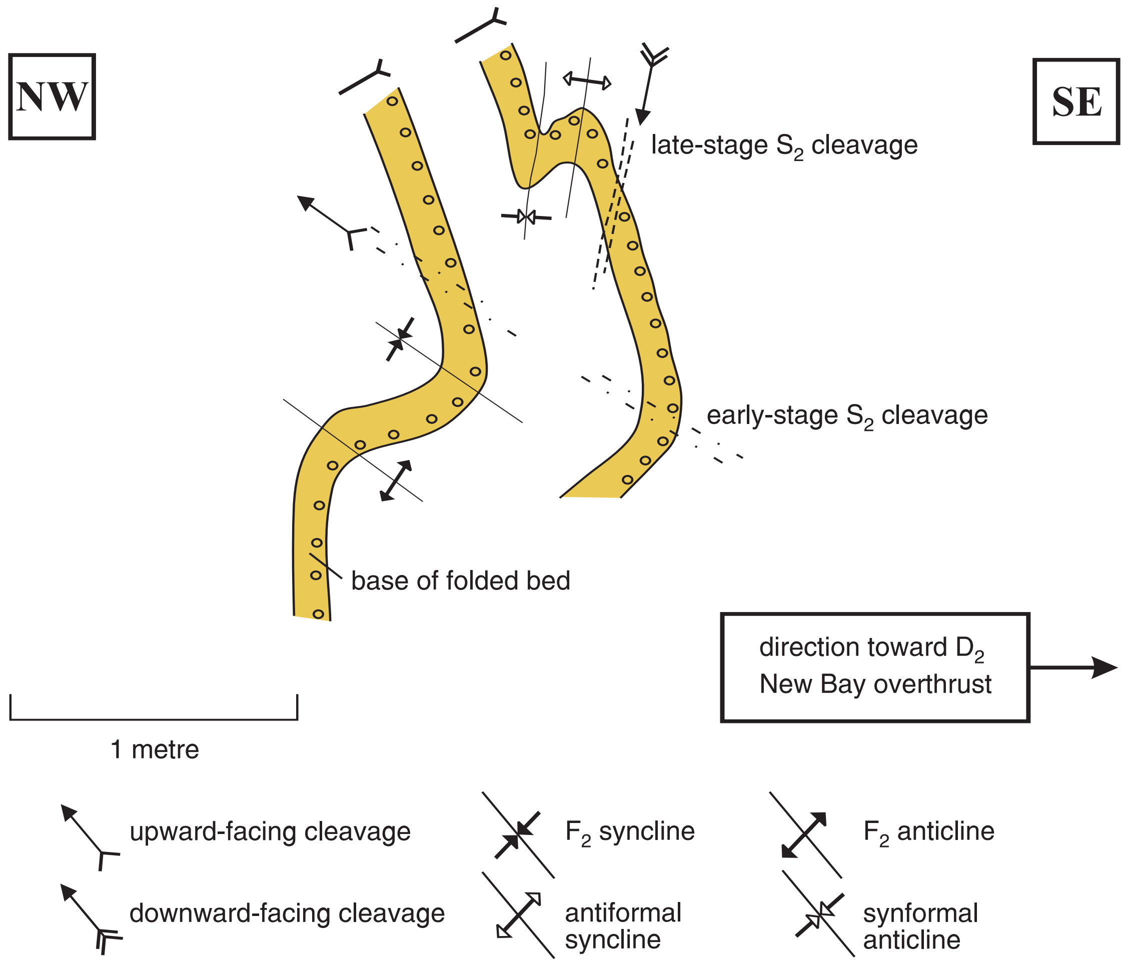


Figure 38. *Schematic cross section illustrating geometrical relationships of multiple generations of S_2 cleavage and antivergent F_2 folds in the Point Leamington Formation along the southeast shore front of Point Leamington Harbour.*

F_2 minor folds defined not only by stratification planes but also small D_1 thrust faults. At exposure [E618750 N5479550] in Little Northwest Arm, the hanging wall sequence of the Point Leamington Formation above such a D_1 thrust is right-way-up; whereas, upside-down strata from the Point Leamington Formation comprise most of the footwall sequence. Therefore, some of the D_3-overprinted F_2 folds in the lower of the two D_1 thrust plates carrying the Point Leamington Formation are antiformal synclines and synformal anticlines (Figure 39a).

Steeply dipping D_3 structures are seen in the same succession of Late Ordovician strata at exposure [E620150 N5480750] in Little Northwest Arm. There, the beds are mainly upside-down in the footwall sequence of a southeast-dipping D_2 reverse fault (Figure 39b). F_3 open folds, S_3 crenulation cleavage and centimetric scale D_3 back thrusts can be traced across a small D_2 fault slice of detached Lawrence Harbour Formation and pass into a structurally underlying tract of the Point Leamington Formation (Figure 39b). In many places, this steeply northwest-dipping S_3 cleavage is observed to face structurally downward.

In central Notre Dame Bay, both the gently inclined and steeply inclined groups of D_3 minor structures are intimately associated with the southeast-dipping D_2 fault zones. Many of the imbricate D_2 reverse faults in these zones record a sinistral-oblique displacement that occurred late in the regional F_2 folding event (*see* Examples of D_2 Structures for the relative timing of the S_2 slaty cleavage, the L_2 extension lineation, the F_2 folds and the D_2 reverse faults). Northwest-dipping D_3 back thrusts, which locally preserve evidence of dextral oblique-slip movement, terminate F_3 (or late-stage F_2) trains of S-shaped folds defined by the surfaces of the regional D_2 reverse faults. Such structures are illustrated in three dimensions in Figure 40. As constructed, the lower D_2 structural plate of the originally planar D_2 sinistral-oblique thrust is also the footwall sequence of the D_3 back thrust.

Interpretation of D_3 Structures

During D_2 sinistral oblique-slip thrusting, the hanging wall plates of the southeast-dipping D_2 thrusts in Little Northwest Arm moved northward and upward away from

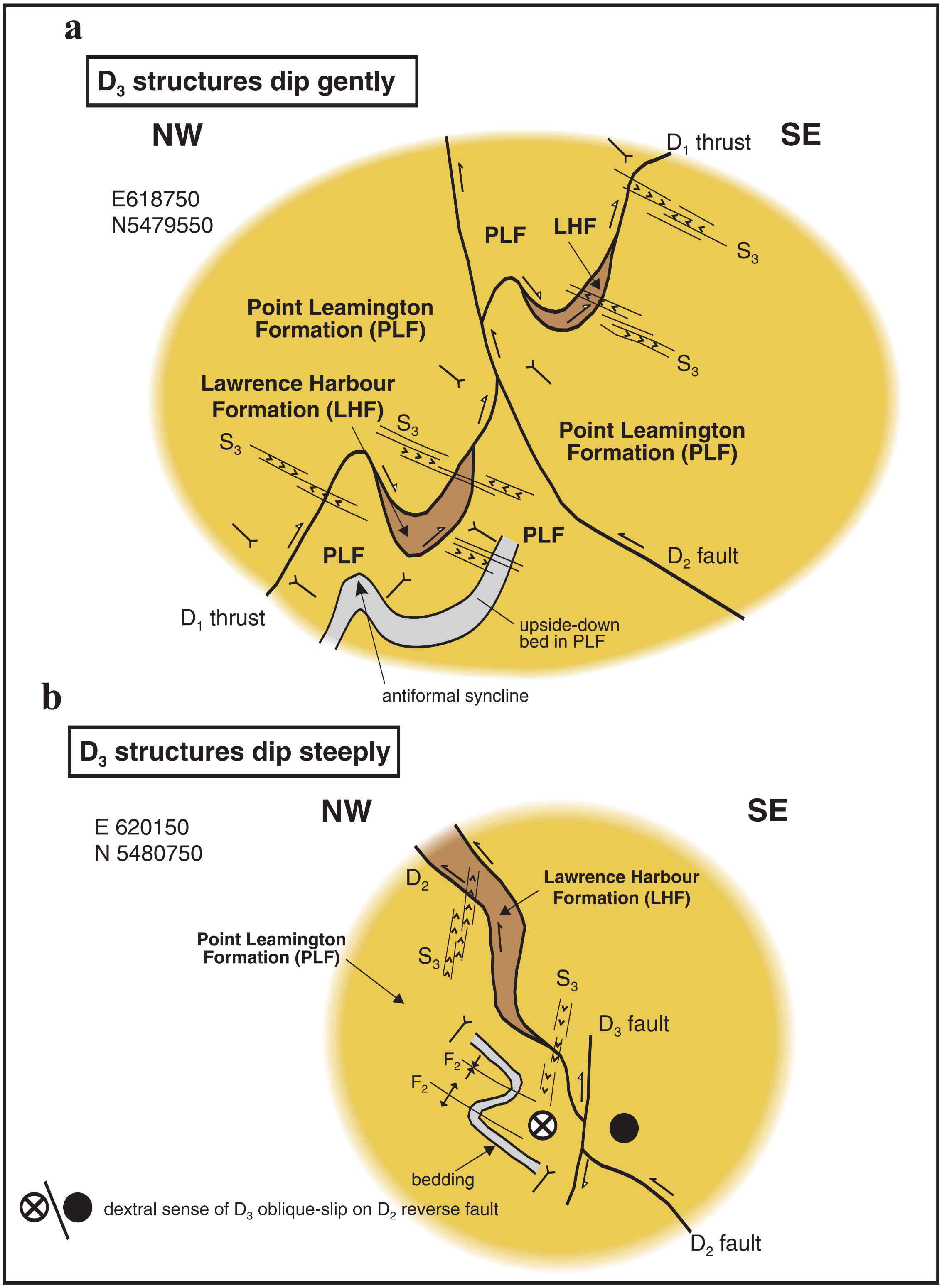


Figure 39. *Cross-sections of two types of D_3 minor structures, viewed looking northeast, in the Upper Ordovician Point Leamington Formation along the northern coastline of Little Northwest Arm. A) Gently southeast-dipping S_3 crenulation cleavage overprints F_2 folds of dominantly northwest-dipping D_1 imbricate thrust faults. Exposure [E618750 N5479550] is located in the hanging wall of the steeply southeast-dipping D_2 Shoal Cove Fault (Figure 25). B) Vertical to steeply northwest-dipping S_3 crenulation cleavage and related S-shaped F_3 open folds overprint a southeast-dipping D_2 reverse fault. Exposure [E620150 N5480750] is located in the footwall of the moderately southeast-dipping D_2 Cold Harbour Fault (Figure 25).*

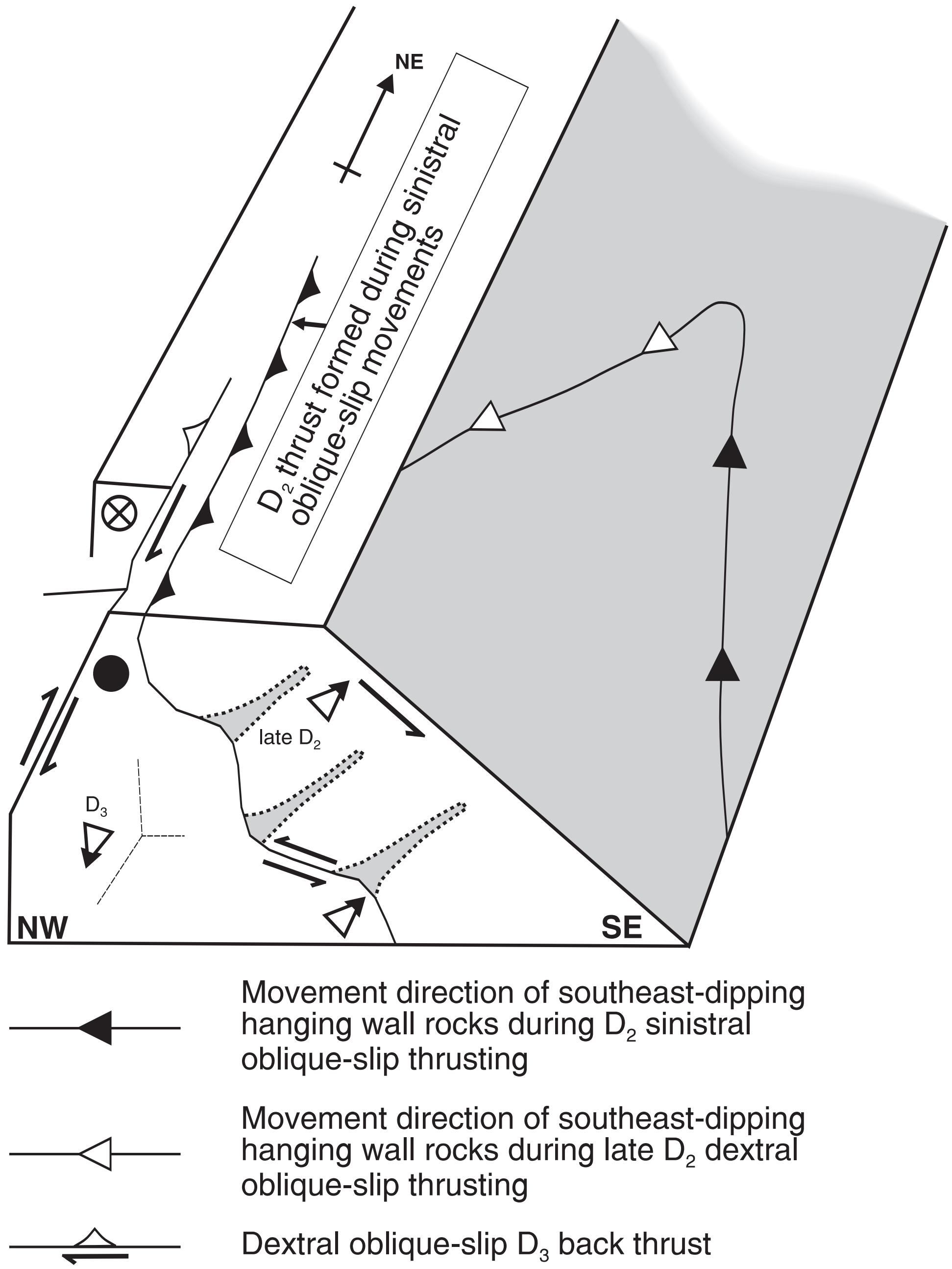


Figure 40. *Block diagram illustrating an interpretation of the localized D_3 back thrusting and F_3 open folding of the south-east-dipping D_2 reverse faults in central Notre Dame Bay. See text for discussion.*

the viewer of Figure 40; whereas, the footwall plates moved southward and downward toward the viewer of Figure 40. However, a late D_2 dextral-oblique slip collision made the footwall sequence beneath original D_2 sinistral-oblique thrusts move in a new direction (downward and away from the viewer of Figure 40). At the same time, the upthrusted hanging-wall sequences ceased to move northward and were instead translated southward. This kinematic switch is portrayed on the shaded surface of the structurally higher D_2 fault as a syn-thrusting change in the movement direction of the southeast-dipping hanging wall plate. Continued southeast-over-northwest translation during dextral shear buckled the sinistral-oblique D_2 thrust faults and created F_3 folds whose axes lay parallel to the late D_2 tectonic movement direction. During D_3 dextral oblique-slip movements, the footwall of the D_3 back thrust moved downward but toward the viewer, as originally did the lower plate of each sinistral oblique-slip D_2 thrust.

Dynamic Analysis

D_3 deformation formed discrete belts of steep- and flat-lying structures, herein referred to as S_3 steep belts and S_3 flat belts. Structures in the southeast-dipping thrust sheet of the Point Leamington Formation below the Cold Harbour Fault (Figure 25) provide an insight into how the steep and flat belts of S_3 cleavage formed (Figure 41). At the end of the D_2 episode of sinistral-oblique shear, this section of the Point Leamington Formation was disposed on the right-side-up and inverted limbs of a northwesterly-overturned fold, carried vertical to moderately southeast-dipping slaty cleavage and was offset by southeast-dipping D_2 reverse faults (Stage I in Figure 41). Like most northeast-trending tracts of rocks illustrating penetrative D_3 deformation, F_2 and F_3 structures in this part of Little Northwest Arm also have opposing senses of fold vergence.

Steeply dipping D_3 structures formed on the inverted limb of the overturned F_2 fold in this part of the Point Leamington Formation; whereas, gently dipping D_3 structures were preferentially developed on the right-side-up limb of this synclinal nappe. During late D_2 and D_3 dextral rotation of the entire fold nappe, dextral-oblique reverse movements occurred on the D_2 fault planes that originally bounded the structure (Stage II in Figure 41). Within the internal part of the nappe, shear couples are postulated to have been active along bedding planes and S_2 cleavage surfaces, where both dipped southeastward. As the effect of the shear couples was most pronounced where all these features were coplanar, the immediate hanging wall and footwall sequences of the southeast-dipping D_2 thrusts illustrate the most complex D_3 structures.

During D_3 regional deformation, reverse faulting occurred along reactivated D_2 fault surfaces. Although sinistral-oblique offset had ceased, D_2 tectonic panels of stratified rocks continued to be transported northwestward during D_3 deformation. Where reactivated D_2 reverse faults in Little Northwest Arm were affected by a dextral-oblique D_3 shear, the F_2 synclinal nappe below the D_2 Cold Harbour Fault was tightened and rotated clockwise. In the lower portion of the F_2 nappe, a D_3 dextral shear couple facilitated the development of gently inclined, Z-shaped F_3 folds, which deformed the S_2 slaty cleavage, further flattened the S-shaped F_2 folds of bedding and made the D_2 bounding fault dip even more gentler. In the upper portion of the F_2 nappe, the D_3 dextral shear couple facilitated the development of steeply inclined, S-shaped F_3 folds, which deformed the S_2 slaty cleavage, interfered with the Z-shaped F_2 folds of bedding and made the D_2 bounding fault dip more steeply (Stage II in Figure 41). The distortion of S_2 slaty cleavage and the concomitant development of S_3 crenulation cleavage are the best indicators of slip related to this episode of convergence, which was oblique to the regional S_2 foliation.

In central Notre Dame Bay, regional D_3 deformation may have been triggered by the gravitational collapse of a vertical stack of fold nappes and overthickened thrust sheets. This could have produced the widespread gently dipping D_3 structures, although it is unusual that conjugate sets of S_3 cleavage did not form. By the end of the D_3 episode of dextral-oblique shearing, northwest-dipping back thrusts had formed (Stage III in Figure 41), apparently at the expense of a southeast-dipping conjugate set of D_3 faults.

D_4 Deformation

Northwest-trending F_4 minor folds and weak S_4 crenulation cleavage are typically developed near northwest-trending D_4 strike-slip faults (O'Brien, 1993b). Sporadically present within thrust sheets and near major fault zones, D_4 structures are most commonly observed to overprint northeast-trending F_2 folds in areas of complex D_1–D_2 interference.

Open northwest-trending flexures of bedding planes and cleavage surfaces are typically observed adjacent to the vertical D_4 faults. In most places, the northeast-trending S_1 or S_2 cleavage displays a normal sense of drag folding in the vicinity of these fault structures (Figure 42). However, in some locations, rather than inducing tectonic movements across the northeast-trending cleavage, D_4 strike-slip faults caused a lateral shear along those bedding planes which had a northwest strike prior to faulting (Figure 42).

Open-to-close F_4 folds with subvertical axial surfaces plunge steeply to gently northwest or southeast. Since the plunge of F_4 structures was mainly controlled by pre-existing variations in the bedding dip or the inclination of S_1 or S_2 foliation, the structural facing direction of axial planar S_4 cleavage on bedding is highly variable.

The predominant effect of D_4 regional deformation was to locally reorient the D_2 structures of northeast-trending domains into parallelism with the high-strain D_1 structures of adjacent northwest-trending domains (Figures 42 and 43). This was mainly achieved by progressive sinistral displacement of D_2 tectonic panels near the trace of northwest-trending D_4 faults; D_4 reactivation of D_1 fault zones has not been recognized in the area surveyed. Where northerly- or northwesterly-trending F_4 major folds developed long limbs and a recognizable sense of fold asymmetry, they outline Z-shaped megakinks situated between northwest-trending D_4 transcurrent faults (Figure 43). Such megakinks are commonly located on the long limbs of the oroclinal Z-shaped F_2 folds.

Interpretation of D_4 Deformation

All D_4 structures can be interpreted as components of a dextral riedel shear system whose principal shear planes were generally oriented along the northeast trend of the orogen (e.g., de Roo and van Staal, 1994; van Staal and de Roo, 1995), but also include regions where the foldbelt locally trends westward (Dube and Lauziere, 1996). In this interpretation, the northwest-trending D_4 faults are sinistral antithetic (R') shears contemporaneous with east northeast-

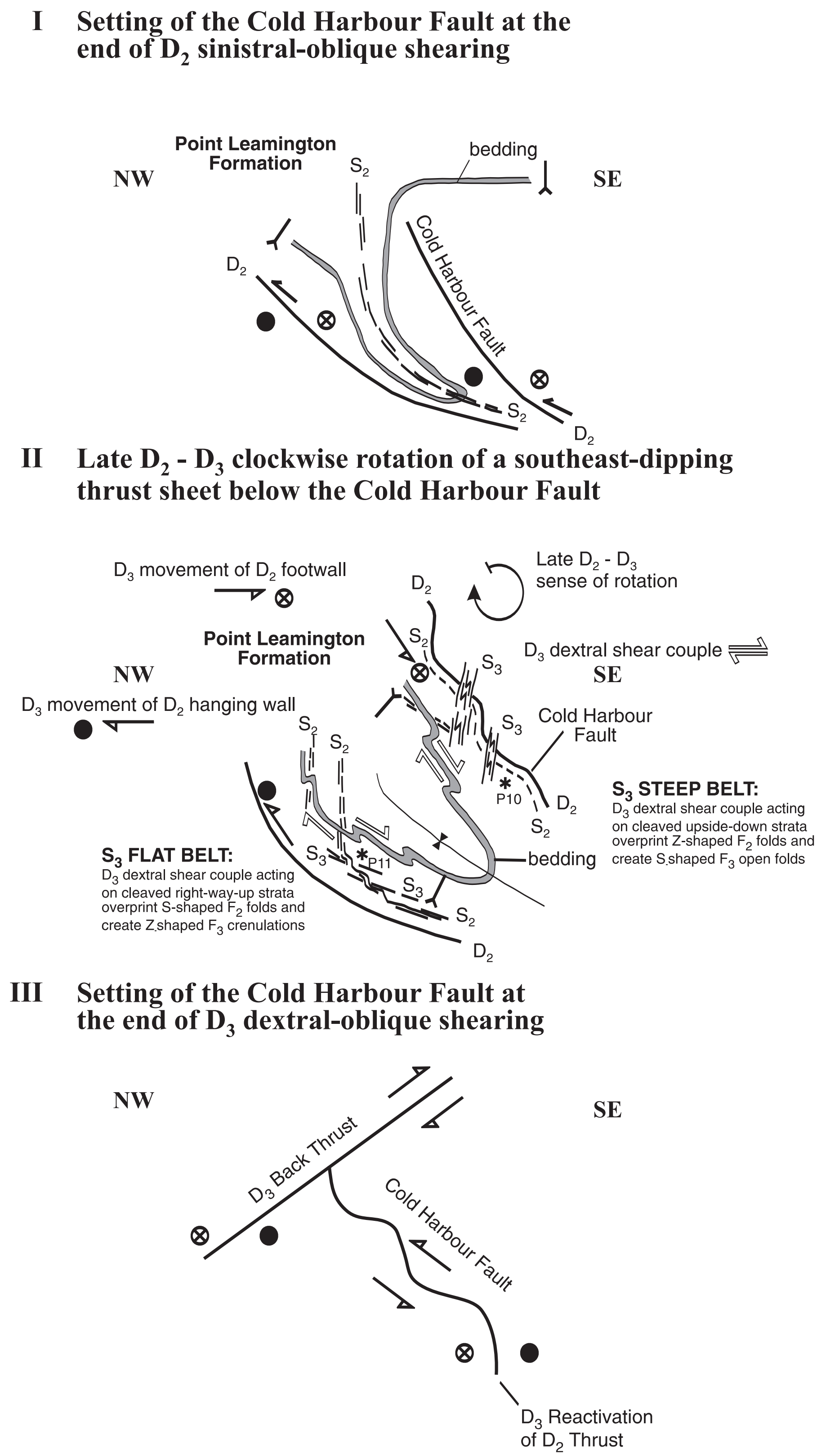


Figure 41. *Three cross sections, viewed looking northeastward, depicting the proposed kinematic evolution of the steeply dipping and the gently dipping D_3 structures. Note the sequential modification of the D_2 structures within a southeast-dipping D_2 thrust sheet. P10* and P11* refer to, and structurally locate, Plates 10 and 11 in O'Brien (1993b).*

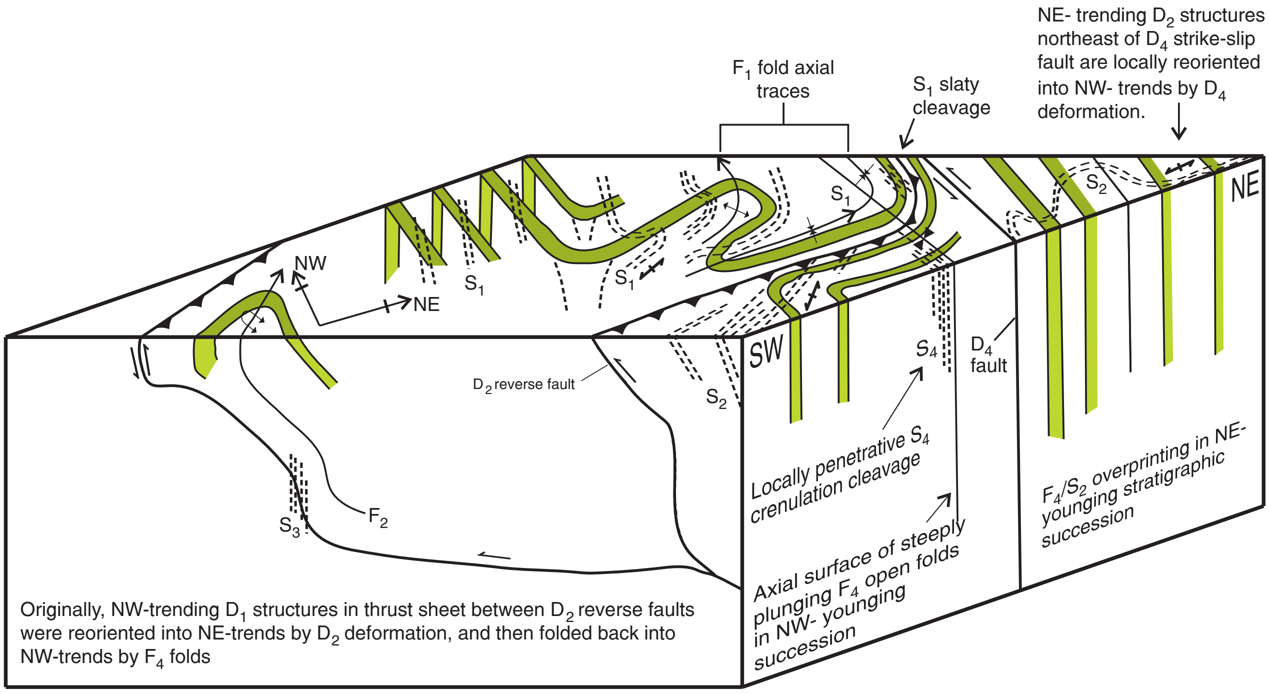


Figure 42. *Block diagram, viewed obliquely northwestward, of characteristic D_1 through D_4 minor structures adjacent to a D_4 fault in the upper Brook Harbour member of the New Bay Formation on the western coast of the Fortune Harbour peninsula. Note that the illustrated northwest-trending vertical fault is a sinistrally transcurrent D_4 structure. The diagram southwest of the D_4 fault is drawn from relationships seen at Exposure [E626600 N5473400] in Indian Cove; whereas, the rocks depicted on the northeast side of the D_4 fault in this diagram occur along strike at Exposure [E624800 N5474700] north of Indian Cove Point.*

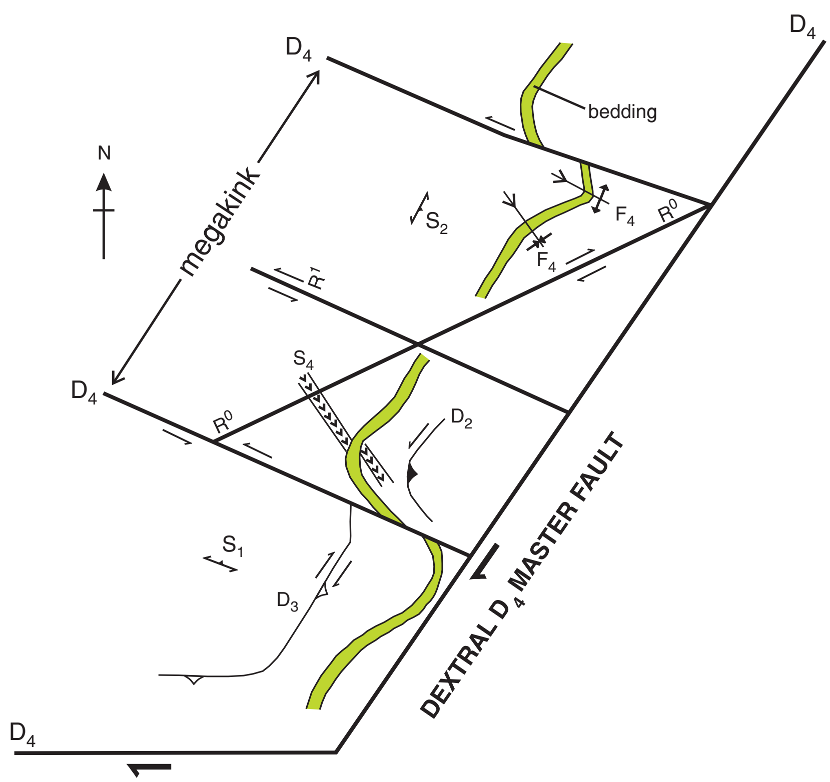


Figure 43. *Simplified structural map showing the typical orientations of F_4 folds, S_4 cleavage and D_4 faults in the area surveyed. Note that these northwest-trending structures can be interpreted as components of a dextral riedel shear system with northeast-trending master faults; observable on numerous scales.*

trending, dextral, synthetic (R) shears (Figure 43). Strata were shortened between these faults to create the F_4 folds, the S_4 crenulation cleavage and the D_4 megakinks.

Prior to emplacement of the posttectonic plutons of the Hodges Hill and Loon Bay batholiths, northeast-trending dextral master faults, such as the ancestral Northern Arm Fault, were still governed by the strongly oblique convergence of the early Silurian rocks in the Exploits Subzone. The east northeast-trending D_4 faults facilitated brittle–ductile displacements sub-parallel to the trend of the orogen by imposing a dextral shear at a low acute angle to the subvertical S_2 cleavage (Figure 43). This permitted the lateral D_4 ejection of previously-stacked thrust sheets and fold nappes near the ancestral Northern Arm Fault under a kinematic regime fundamentally different than the one which controlled the D_2 episode of tectonic escape.

PLUTONIC ROCKS OF CENTRAL NOTRE DAME BAY

In the study area, plutonic rocks comprise the main intrusive phases of major batholiths, occur as regionally developed suites of minor intrusions, and form the major components of several large igneous complexes. The individual relationships of plutonic rocks to Siluro-Devonian structures permit assignment of intrusions to pre-, syn-, and posttectonic suites. These magmatic suites, in particular the pretectonic intrusions, relate to several distinct Ordovician and Silurian phases of Dunnage Zone development. In central Notre Dame Bay, it is common for a map unit of crystalline igneous rocks to have intrusive, faulted or nonconformable boundaries and, depending on the age of the pluton and the age range of adjacent stratified units, to display more than one type of geological contact.

PRETECTONIC MAJOR INTRUSIONS

Phillips Head Igneous Complex

The Phillips Head igneous complex is an informal unit composed of pyroxene-porphyritic vesicular diorite, various suites of mafic hypabyssal intrusions and enclaves of amygdaloidal pillow lava and pillow breccia (Plate 38; Figure 44). All of these rock types are texturally similar and, for the most part, are unique to the igneous complex. Most hypabyssal intrusions in the complex are seen to crosscut vesicular diorite and pillowed basalt; however, intrusive contacts between diorite and basalt have not been directly observed.

Regional thrust faults or brittle structures related to the Northern Arm Fault bound the Phillips Head igneous complex and separate it from the Exploits Group, the Wild Bight Group, the Shoal Arm Formation and the Botwood Group (Figures 15 and 44). The unit was previously correlated with mafic plutonic and hypabysal phases of the Hodges Hill batholith and was deemed to be in intrusive contact with the Point Leamington Formation of the Badger Group (Kean *et al*., 1981).

Lithology

The most distinctive features of the large diorite body in the Phillips Head igneous complex are its ubiquitous vesicularity and extensive brecciation. Coarsely porphyritic, amygdaloidal diorite fragments are surrounded by an igneous matrix of less coarse, vesicular, diorite porphyry. In many locations, the early crystallized diorite fragments are affected by a widespread phase of saussuritization. This alteration predated inclusion of the saussuritized plutonic fragments into the magma batch represented by the fresh diorite porphyry matrix. It suggests that the brecciation and

Plate 38. *Monolithic volcanic breccia from the Phillips Head igneous complex, composed of highly vesicular blocks of Wild Bight-type basalt, is intruded by chilled mafic dykes emanating from a diorite screen on the viewer's left.*

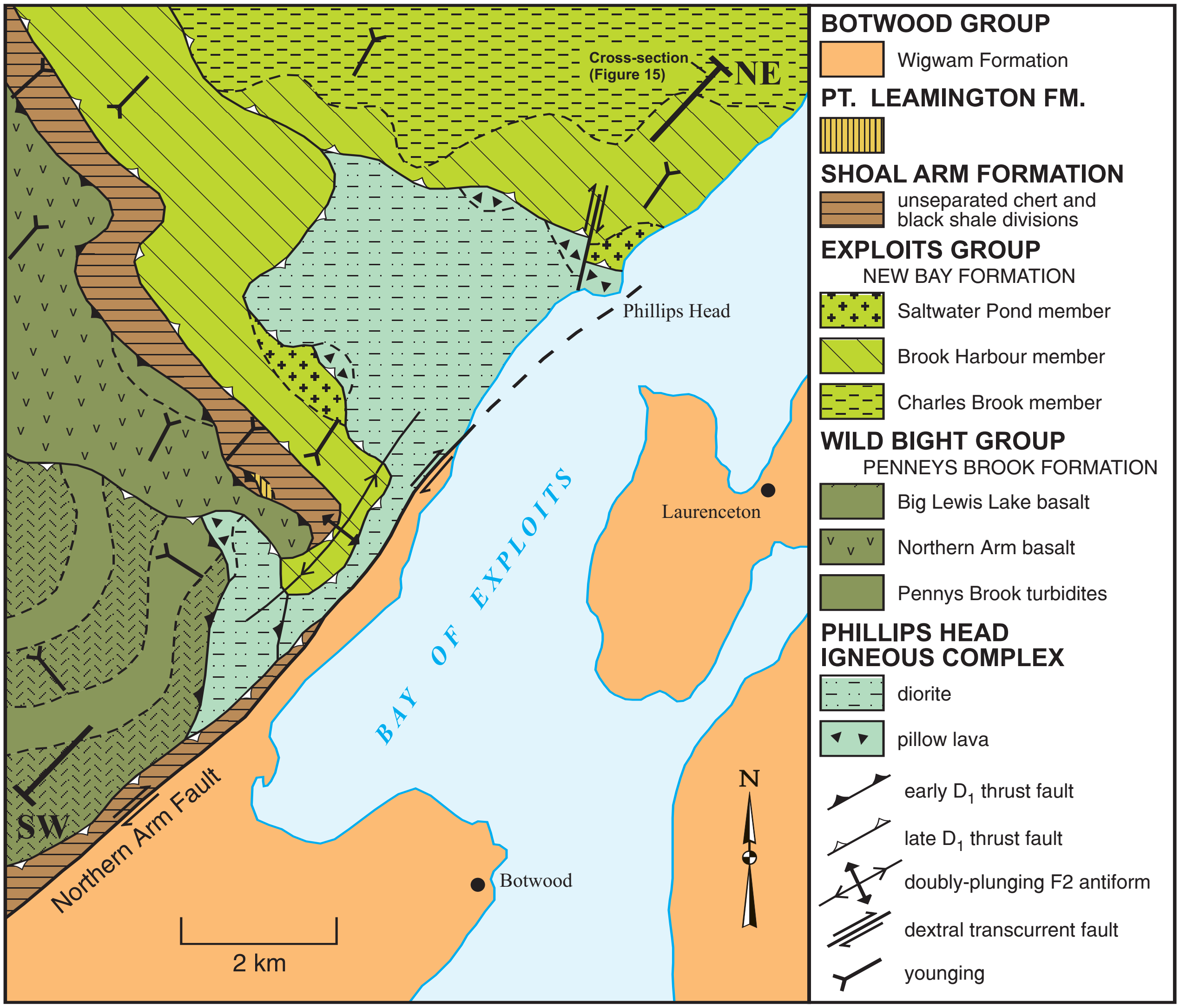


Figure 44. *Simplified geological map of the Phillips Head igneous complex showing the relative distribution of constituent intrusive and extrusive rocks. Also depicted are the complex's external relationships with the New Bay Formation of the Exploits Group, the Pennys Brook Formation of the Wild Bight Group, the Shoal Arm Formation, and the Wigwam Formation of the Botwood Group. The line of section for Figure 15 is also illustrated.*

alteration of Phillips Head diorite occurred within a high-level magma chamber, which periodically lost some of its volatile content during crystallization.

Unique features of the Phillips Head diorite are the lack of any satellite intrusions emanating from the main diorite pluton and the absence of any internal granitoid or pegmatitic phases. Other field characteristics that distinquish Phillips Head diorite from Hodges Hill gabbro are the lack of any evidence of stoping of adjacent stratified units, the lack of assimilation of pillow lava enclaves, and the lack of a thermal metamorphic aureole around any part of the plutonic body.

The minor intrusions in the Phillips Head igneous complex are unique amongst the hypabyssal rocks of central Notre Dame Bay for several reasons. First, intrusive breccias are locally developed at the margins of some of the larger dykes in the complex and tuffisite pipes contain comminuted fragments of diorite. The fluidization or gas-streaming textures in the breccias may genetically relate such high-level mafic dykes to the host diorite, which is also characterized by devolatization textures. Second, the abundant aphanitic-, porphyritic- and schlieren-textured mafic dykes within the igneous complex are unclustered and commonly undeformed. In this regard, they contrast with the mesozonal syntectonic dykes in the Ordovician diorites and tonalites of other pretectonic igneous complexes in the study area.

Evidence for a Pretectonic Origin

The external boundaries of the Phillips Head igneous complex are interpreted to be folded D_1 faults which trend northwestward and northeastward (Figure 44). These are displaced by the dextrally transcurrent Northern Arm Fault. Although exposures of D_1 fault surfaces have not been located, these boundaries are considered to be thrust or reverse faults. This is because S_1 cleavage transected and S_2 cleavage overprinted folds of bedding in adjacent stratified units as the regional bulk strain increased toward the border of the igneous complex. Also because strata immediately northeast of the Phillips Head diorite were upside-down prior to the development of the northeast- trending D_2 minor structures that deform these rocks and the margin of the igneous complex. A northeast-trending, doubly plunging F_2 antiform located farther southwest is an analogous major structure which also folds a boundary fault of the Phillips Head igneous complex (Figures 15 and 44). In addition, structurally-controlled vein arrays deformed by gently-dipping S_2 or S_3 crenulation cleavage illustrate thrust-sense kinematic indicators along some of the external boundaries of the complex.

Age and Correlation

The emplacement age of the diorite and the eruptive age of the basalt in the Phillips Head igneous complex are unknown. Although it is possible that the intrusive and extrusive rocks are consanguineous, it is also probable that the complex's plutonic rocks are more closely related to the gabbro laccoliths found throughout the Exploits and Wild Bight groups than to the Phillips Head volcanic strata (O'Brien *et al*., 1997). Based on regional considerations, the Phillips Head igneous complex is probably mid Ordovician in age.

Volcanic Rocks

The Phillips Head pillow lavas are interpreted as roof pendants of right-way-up volcanic rocks lying above the high-level Phillips Head diorite. On the basis of their lithology, structure and geochemistry (O'Brien *et al*., 1994; Hughes and O'Brien, 1994), these pillow lavas are similar to the calc-alkaline basalts in the lower Pennys Brook Formation of the upper Wild Bight Group. They are generally dissimilar to the within-plate basalts of the Lawrence Head Formation of the upper Exploits Group, which is tectonically excised along the fault contact separating the inverted panel of the New Bay Formation of the Exploits Group from the structurally underlying Phillips Head basalt and diorite (Figure 15). The Phillips Head pillow lavas are geochemically similar to the basalt olistoliths in the Saltwater Pond Member of the New Bay Formation of the Exploits Group (Hughes and O'Brien, 1994).

Plutonic Rocks

A posttectonic origin for the Phillips Head diorite and a correlation with the nearby Hodges Hill intrusive complex is negated, if one accepts the geological evidence for folded D_1 faults along the complex's boundaries. However, the Phillips Head diorite may be possibly related to some of the pretectonic mafic intrusive rocks in other parts of central Notre Dame Bay. Such pre-Salinic plutonic rocks include the sheared and undeformed diorites in the Early Ordovician part of the South Lake Igneous Complex, the folded diorite sills hosted by the Early to Middle Ordovician Exploits and Wild Bight groups, the regionally foliated diorite sheets intruding the Late Ordovician to Early Silurian Badger Group, and the variably deformed diorites emplaced into the basalts of the Middle to Late Silurian Botwood Group.

Geochemistry

Preliminary geochemical analyses of samples of the Phillips Head diorite and a crosscutting Phillips Head diabase show that these intrusive rocks are LREE-enriched ([La/Yb]n= 5.85 and 3.93, respectively). By way of comparison, Phillips Head pillow lavas are LREE-depleted ([La/Yb]n=0.678) and extremely subalkalic (Nb/Y=0.068), and have been interpreted as arc-related basalts with clear positive Th and negative Nb anomalies (*see* LREE-plots in Hughes and O'Brien, 1994). Thus, despite some textural similarities, intrusions within the complex are petrochemically distinct from the enclaves of constituent volcanic rocks. However, Phillips Head extrusive rocks are similar to flows in the Northern Arm Basalt of the Wild Bight Group (Swinden *et al*., 1990; MacLachlan, 1998).

Tectonic discrimination diagrams involving the less incompatible elements Ti and Y indicate a non-arc affinity (ocean floor basalt) for the transitional subalkalic to alkalic diorite (Ti/V= 23.8; Nb/Y= 0.74) and the slightly subalkalic diabase (Ti/V= 48; Nb/Y= 0.27) of the Phillips Head igneous complex. An extended REE-plot of the Phillips Head diorite shows a strong positive Th and weak negative Nb anomaly relative to La and adjacent LREE elements. In the diabase, however, normalized concentrations of Th and Nb are both lower than La, although Th is slightly elevated relative to Nb. Accordingly, in discrimination diagrams involving Th (e.g., Gill, 1981), the Phillips Head diorite plots within the field of high-K orogenic andesites (La/Th= 4.381). In contrast, the Phillips Head diabase plots within the N-MORB field (La/Th= 13.870).

LREE-enriched rocks similar to the intrusions in the Phillips Head igneous complex occur in the tectonically adjacent Exploits Group. Examples include the Llanvirn-Llandeilo high-K calc-alkalic gabbros on Thwart Island (Ti/V=154, Nb/Y=0.23, La/Th= 6.328, [La/Yb]n= 2.99) and some tholeiitic pillow lavas in the section of the late Arenig Lawrence Head Formation at nearby Purbecks Cove (O'Brien *et al*., 1997). In tectonic discrimination diagrams, such Exploits Group rocks plot, variably, in non-arc (Ti/V) or arc (Nb-Hf /Y) fields, much like the intrusions in the Phillips Head igneous complex.

Structural Setting

Regionally, the Phillips Head igneous complex is situated in a southeast-plunging thrust stack which also carries fault-duplicated parts of the Exploits and Wild Bight groups (Figure 15). Between the northeastern and southwestern segments of this complex, a southeasterly elongated F_1 dome is cross folded by a periclinal F_2 antiform to produce a map-scale interference pattern and a regional fold culmination of the thrust along the base of the Phillips Head diorite (Figure 44). This antiformal culmination provides a window into thrust sheets that comprise the structural footwall sequence lying beneath the Phillips Head igneous complex.

On passing from the northwestern termination of the northern segment of the Phillips Head igneous complex, the northeasternmost boundary fault is strata-parallel and lies completely within the New Bay Formation of the Exploits Group (Figure 44). However, along strike farther northwest, the same structure displaces a northwest-plunging syncline in the Point Leamington Formation and delimits the southernmost outcrop of that formation (Williams, S.H. and O'Brien, 1994). This fault-modified syncline culminates in early and mid Ordovician units of the Exploits Group (Figure 15) and plunges regionally southeastward near the northern segment of the Phillips Head igneous complex.

On passing from the northwestern termination of the southern segment of the Phillips Head igneous complex, the southeasternmost boundary fault merges with a fault structure that lies completely within the Wild Bight Group (Figure 44). Farther to the northwest, the same structure displaces a southeast plunging syncline in the Pennys Brook Formation (Figure 15).

Thus, the thrust sheet containing the Phillips Head diorite is structurally overlain and underlain by the strata of the New Bay Formation and the Pennys Brook Formation. The structural triangle zone in the Shoal Arm Formation (Figure 30) is interpreted to lie beneath these formations in the footwall sequence of the Phillips Head diorite at a deeper level in the folded imbricate stack of plutonic and stratified rock units (Figure 15).

Phillips Head Cataclasites

Cataclastic deformation of the Phillips Head igneous complex along the presumed Devono-Carboniferous Northern Arm Fault produced random fabric fault gouge and cohesive crush breccia. These features are also observed to a lesser degree farther south in the posttectonic Northern Arm pluton (Dean, 1978) of the Hodges Hill intrusive complex. Within 30 m of the fault trace, there is excellent evidence of progressive grain-size reduction of the Phillips Head diorite, which was comminuted in the matrix of the crush breccia. Some of this tectonized plutonic material has been previously mapped as massive Point Leamington Formation wacke (e.g., Kean *et al*., 1981).

Well-foliated zones of dominantly northeast-trending, quartz-veined chlorite schist, apparently derived from Phillips Head diorite and diabase, are also present in the vicinity of the random fabric gouges. Small ductile shear zones hosted by highly altered basalt and interstitial chert are also present within this fault zone. In exposures of the Northern Arm Fault east of Muddy Hole Brook, silicified Phillips Head diorite is juxtaposed against a tectonic lens of brecciated and recrystallized Shoal Arm Formation (Figure 44).

In the vicinity of this vertical strike-slip fault zone (Kusky, 1996; Kusky *et al*., 1987), shear bands dip gently northwestward within the Phillips Head igneous complex and the adjacent Botwood and Wild Bight groups. These brittle–ductile shears record an oblique north-over-south sense of dextral overthrusting and are geometrically similar to the structures shared by the Hodges Hill granite and the Wigwam Formation of the Botwood Group (Figure 17). They probably relate to Devono-Carboniferous deformation of the Phillips Head igneous complex, although they could be a relic of regional Siluro-Devonian deformation.

South Lake Igneous Complex

The South Lake Igneous Complex is composed of gabbro, diorite and tonalite bodies which host several swarms of mafic dykes (O'Brien, 1992; Figure 45). All of these plutonic and hypabyssal rocks have been variably deformed and metamorphosed, and some have been affected by multiple dynamothermal events. Within the igneous complex, South Lake plutons display mostly intrusive boundaries. In contrast, the external boundaries separating the complex from adjacent stratified rock units are mostly folded D_1 faults. Large enclaves of volcanic rocks have not been recognized in the South Lake Igneous Complex, although small accidental xenoliths of highly deformed tectonites are present in several of the constituent plutons.

Historically, the South Lake Igneous Complex has been viewed as comprising distinctive tracts of Cambro-Ordovician ophiolitic and magmatic arc rocks (Dean, 1977; Lorenz and Fountain, 1982). Preliminary U/Pb zircon dating indicated earliest Ordovician crystallization ages for tonalite bodies on the west and northeast shore of South Lake (G.R. Dunning, unpublished data, 1992); more recently, a metagabbro and several crosscutting tonalite and diorite bodies have also yielded similar absolute ages (MacLachlan, 1998). Thus, the South Lake suite of deformed plutonic rocks are considerably older than the adjacent formations of stratified rocks, three of which are known to range, collectively, from the middle to the Late Ordovician (Figure 45).

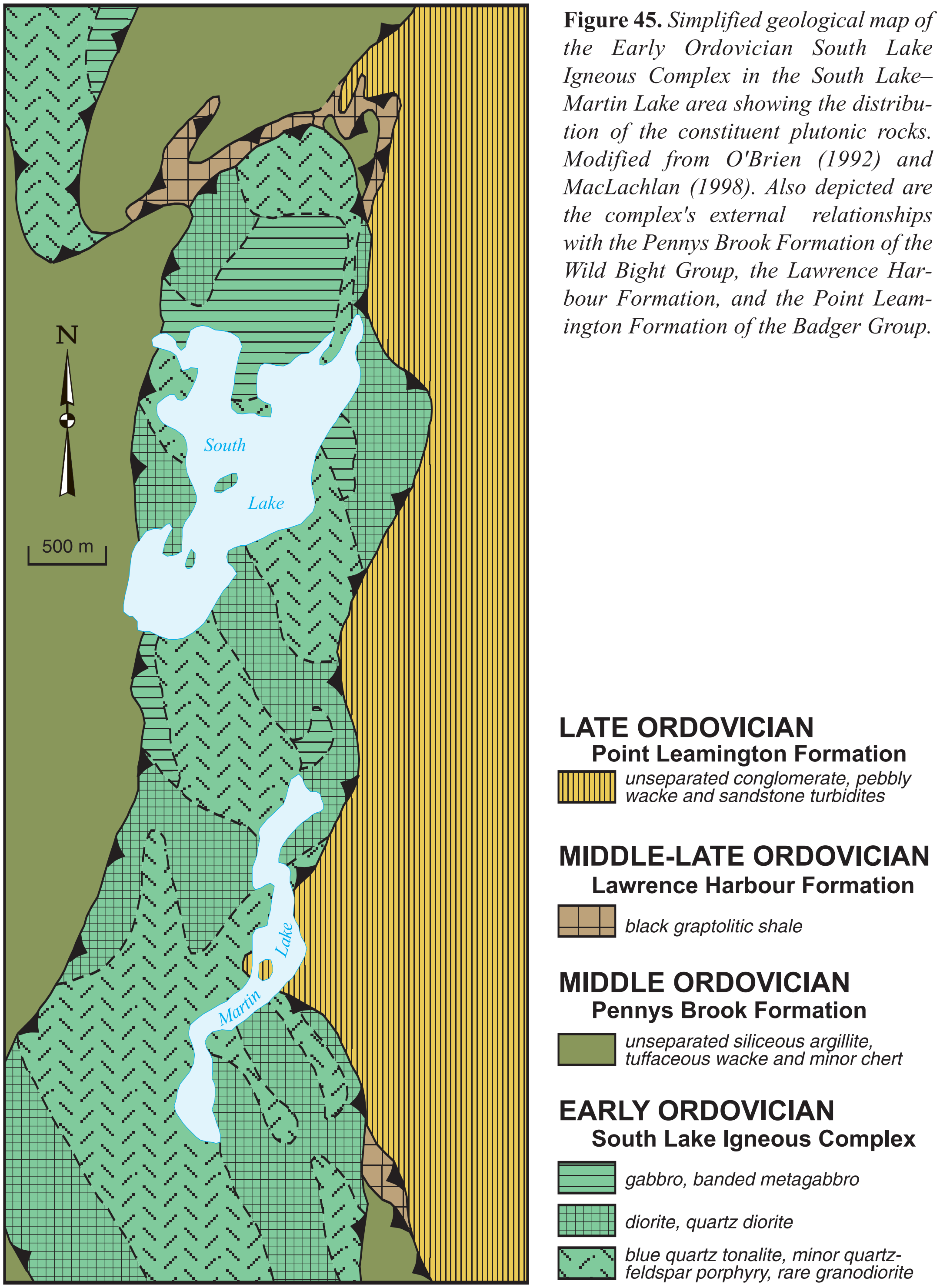


Figure 45. *Simplified geological map of the Early Ordovician South Lake Igneous Complex in the South Lake–Martin Lake area showing the distribution of the constituent plutonic rocks. Modified from O'Brien (1992) and MacLachlan (1998). Also depicted are the complex's external relationships with the Pennys Brook Formation of the Wild Bight Group, the Lawrence Harbour Formation, and the Point Leamington Formation of the Badger Group.*

Constituent Units

The oldest exposed part of the South Lake Igneous Complex, originally referred to as the South Lake Ophiolite (Dean 1977), is best preserved in the northern part of the complex (Figure 45). There, relatively small tracts of metagabbro are intruded and totally surrounded by dated early Ordovician plutonic bodies. In a few places, metagabbro is faulted directly against the Pennys Brook Formation of the Wild Bight Group. Ultramafic rocks have not been recognized within the main body of the South Lake Igneous Complex.

Metagabbro

The oldest unit in the igneous complex consists of northeast- to east-trending zones of cumulate-layered, schistose and flaser-banded gabbro. A host-rock foliation commonly lies parallel to cumulate layering and is observed to be axial planar to isoclinal folds of *lit-par-lit* leucogabbro veins that crosscut compositional banding. Thus, field relationships of *lit-par-lit* leucogabbro veins to variably sheared and foliated metagabbro indicate that intrusion of younger phases of this pluton accompanied deformation and metamorphism of older phases. The leuocogabbro veins are thought to occupy a series of branched, asymmetrically-forked shear fractures that developed during transient brittle deformation in ductile shear zones hosted by the metagabbro pluton (*see* Plate 1 of O'Brien, 1992a).

Sheeted mafic dykes were emplaced into small northwest- to west-trending shear zones that crosscut eastnortheast-trending foliation, flaser banding and igneous layering in the metagabbro (Plate 39). These dykes comprise the oldest recognizable swarm in the South Lake Igneous Complex and are most abundant within amphibolitized parts of the host pluton. Steeply dipping hornblendite veinlets, which illustrate distinctive en echelon geometric arrangements in highly altered metagabbro, crystallized from late magmatic or metamorphic fluids prior to the intrusion of the sheeted dykes. The hornblendites are, however, interpreted to have filled small-scale tensional structures that abutted strike-slip fractures now marked by the traces of the mafic intrusions (*see* Plate 2 of O'Brien, 1992a).

An internal wall-to-wall foliation is commonly developed in the discordant suite of northwest-trending mafic dykes. Locally, near dyke walls, a coplanar fabric is developed in adjacent country rocks, where it is seen to be axial planar to tectonic folds of primary igneous layers. The apical angle of such minor folds decreases toward the margins of the most strongly foliated dykes. Variably sized rhombohedral blocks of cumulate-layered or flaser-banded gabbro were isolated from each other by the abutment of the main northwest-trending swarm of dykes with a subordinate set of concordant northeast-trending diabases (*see* Plate 3 of O'Brien *et al*., 1992a). In the northern part of the complex, angular enclaves of layered gabbro and sheeted dykes form northwesterly oriented trains of the ophiolitic unit engulfed by foliated diorite and tonalite bodies (*see* Cawood *et al*., 1995 for similar geometric pattern of arc ophiolite relics and arc root plutons).

Northeast-trending protomylonite zones, each several metres wide, rework metagabbro and the originally discordant northwest-trending mafic dykes. Inhomogeneously developed, they represent some of the youngest ductile structures within the oldest part of the South Lake Igneous Complex. These subvertical mylonite zones are older than the diorite plutons of the South Lake Igneous Complex.

Plate 39. *Mafic dykes, which crosscut the flaser banding of the oldest metagabbro unit of the South Lake Igneous Complex, occupied synmagmatic cracks that are seen to have curved tips and en echelon terminations.*

Hornblende Diorite

Diorites and tonalites make up most of the southern part of the South Lake Igneous Complex. They have been purported, on geochemical grounds, to comprise a comagmatic suite of primitive arc-related plutonic rocks (Lorenz and Fountain, 1982; MacLachlan and Dunning, 1998a).

Subvertical bodies of hornblende diorite crop out as septa between tonalite sheets and as screens on the margins of the metagabbro unit. They vary from fine- to coarse-grained, equigranular to schlieren-textured, and commonly contain mafic pegmatites in diffuse pods or cavities. Most bodies of hornblende diorite are unfoliated or weakly schistose; many are fresh and some are locally altered.

Hornblende diorite is host to variably sheared dykes of mafic and felsic composition. These minor intrusions are localized in small northwest-trending trains near those segments of the plutons rich in accidental and cognate xenoliths. They are not, however, present in all of the hornblende diorite bodies in the area. The association of dyke rocks and parallel trains of xenoliths is most common in the northern part of the South Lake Igneous Complex.

Quartz Tonalite

The most lithologically distinctive and physiographically prominent member of the South Lake Igneous Complex is a coarse-grained tonalite rich in blue quartz. Throughout most of the complex, such quartz tonalite bodies are spatially associated with the hornblende diorites. Moreover, tonalite plutons are commonly bordered by zones of composite mafic–felsic dykes.

Coarse grained, quartz-porphyritic tonalite has ophitic intergrowths of much finer plagioclase, pyroxene and hornblende that are interpreted as cognate inclusions and xenocrysts. Small irregular clots of these minerals are gradational with the tonalite matrix; locally ophitic microdiorite forms discrete trains of cognate xenoliths within tonalite. This early crystallized material is very common in all the blue-quartz tonalite plutons and is especially well-developed near swarms of distorted, distended and disaggregated mafic dykes.

Well-foliated but poorly lineated bodies of quartz tonalite are widespread in the northern part of the igneous complex. In places, the northwest-trending subvertical tonalite schistosity is defined by reorientated and neocrystallized ferromagnesian minerals. In contrast, unrecrystallized magmatically zoned feldspars show a primary mineral alignment in other parts of the tonalite plutons. In both deformed and unmetamorphosed states, tonalite bodies are host to the second major swarm of mafic dykes in the South Lake Igneous Complex.

Quartz tonalite is observed to be intruded as a series of anastomosing and tapering sheets. Tonalite screens typically contain variably sized fragments of hornblende diorite; diorite septa are injected by tonalite stringers and overgrown by metasomatic feldspars. Small ductile shear zones observed within some diorite septa are absent in adjacent tonalite screens, possibly implying that the initiation of diorite deformation may have preceeded tonalite intrusion in certain areas.

Accidental xenoliths of mafic and silicic rock types are found in blue-quartz tonalite but they are not as widespread as the entrained cognate inclusions. Most of the accidental xenoliths are highly strained tectonites; some banded migmatitic varieties display complex fold interference patterns. All of the structures seen in the xenoliths predated tonalite intrusion and cannot be younger than Early Ordovician in age. Xenoliths of flaser gabbro and hornblende diorite in quartz tonalite indicate that the tonalite plutons ascended through a tectonically stitched assemblage of the metagabbro and diorite units.

One explanation of the ductile structures preserved in the banded mafic inclusions in tonalite is that they were generated during the early mylonitization of the older metagabbro and sheeted dyke unit. However, the xenoliths of silicic tectonites are not simply related to the early deformation of the ophiolitic tract, as evidence for extensive silicification and replacement of metagabbro is lacking in the region. Such xenoliths are probably exotic tectonites sampled from depth, since the altered felsic volcanic rocks in the Early Ordovician part of the Wild Bight Group were not foliated prior to the intrusion of the South Lake tonalite bodies.

Granodiorite

A large body of quartz-rich granodiorite transitional to microgranodiorite and microtonalite occurs in the northeasternmost part of the South Lake Igneous Complex, where it directly overthrusts the black shales of the Lawrence Harbour Formation (Figure 45). The youngest member of the complex, it may be related to the blue-quartz tonalite by fractionation (MacLachlan, 1998).

In most places, the granodiorite is a medium-grained equigranular rock, which locally becomes porphyritic where the intrusion is chilled against the metagabbro unit. It is rarely foliated, lacks cognate inclusions and contains uncommon mafic dykes. The margin of the granodiorite intrusion north of South Lake is mapped, in part, as transecting the intrusive contact between hornblende diorite and metagabbro and, in other localities, as occupying that boundary. Marginal facies of the granodiorite are fine-grained, quartz porphyritic and feldsparphyric rocks.

On the shore of South Lake, a north-trending unfoliated sheet of strongly jointed granodiorite contains accidental xenoliths of mylonite (Plate 40). Distinctive epidote-rich bands within the mylonite suggest derivation, at least in part, from rocks of mafic composition. Highly sheared and partly silicified, blue quartz-bearing rocks occur as xenoliths in unaltered and unfoliated granodiorite near the northwest margin of this intrusion. Elsewhere, within the main intrusive body, the granodiorite itself is locally saussuritized, sericitized and pyritized. Sulphide-rich alteration zones are also present where small apophyses of granodiorite intrude older members of the igneous complex.

Mafic Dyke Emplacement

Minor mafic intrusions were successively emplaced into the South Lake Igneous Complex during several geologically discrete events. Such dykes were intruded into progressively younger magmatic rocks within the confines of the complex, although the Early Ordovician time range for most of these events is demonstrably narrow (G .R. Dunning, unpublished data, 1992; MacLachlan, 1998). Some pertinent facts which relate to the sequential swarming of South Lake mafic dykes are listed below.

(i) Strongly deformed tracts of metagabbro, which contain undeformed northwest-trending diabase dykes belong-

Plate 40. *Unfoliated quartz diorite from the South Lake Igneous Complex contains xenoliths of foliated and sheared metabasite (beneath the hammer), which is most similar to older rocks in adjacent parts of the complex.*

ing to the oldest recognizable swarm in the complex, are crosscut obliquely by northeast-trending subvertical plutons of hornblende diorite. In other localities, the original intrusive boundaries of steeply dipping bodies of undeformed hornblende diorite trend northwest. There, they appear to truncate vertical zones of northeast-trending protomylonite which had developed in the adjacent reworked tracts of diabase-injected metagabbro.

(ii) Hornblende diorites generally display igneous textures and are isotropic but, in many places, they are well-foliated. In both undeformed and deformed states, they are host to small locally developed swarms of northwest-trending diabase dykes. These intrusions are significantly younger than mafic dykes of similar orientation within the metagabbro unit. In places, northwest-trending, metre-thick sheets of tonalite occupy the intrusive contact between the younger set of diabase dykes and the host hornblende diorite. In other locations, the walls of mafic dyke chambers in hornblende diorite are marked by en echelon trains of centimetric-scale felsic dykelets which are probably related to the adjacent quartz tonalite bodies. None of these minor mafic intrusions are observed in the tectonically adjacent Pennys Brook Formation of the Wild Bight Group.

(iii) Blue-quartz tonalite and granodiorite contain xenoliths of an amphibolitized gabbro that carries crosscutting dykelets of foliated diabase. These xenoliths are found in an area where the tonalite and granodiorite plutons come in direct intrusive contact with a large unit of metagabbro. This, together with the intrusive relationship between a hornblende diorite body and the same metagabbro unit, suggests persistent sampling of the oldest plutonic rocks in the igneous complex during the ascent and emplacement of the younger diorite, tonalite and granodiorite bodies.

(iv) Cognate xenoliths in the complex's coarse-grained blue-quartz tonalites were flattened and elongated prior to being intruded by another generation of mafic dykes. These mafic dykes were subsequently distended and, as a result, partially fragmented within the host tonalite. All of these deformed rock types were then locally attenuated during magmatic flow within small intra-plutonic shear zones. In certain parts of the tonalites, magmatic foliations are defined by the alignment of plagioclase megacrysts and the orientation of mafic cognate xenoliths. The deformation of cognate xenoliths and mafic dykes within tonalite overlapped the deformation of the material backveined from the tonalites into the synplutonic mafic dykes.

(v) Blue-quartz tonalite was also ductilely deformed in the solid state in several areas within the South Lake Igneous Complex. There, tonalite bodies display a crystal-plastic foliation that is best defined by ribboned grains of grey quartz. Undeformed mafic dykes of unknown age are observed to crosscut the ribboned-quartz fabric in metamorphosed and schistose tonalite. Restricted to the shear zones found in the northern part of the South Lake Igneous Complex, they may represent some of the youngest phases of mafic dyke intrusion in the complex.

Summary of Ordovician Events

Ordovician episodes of ductile deformation and several periods of dynamothermal metamorphism were coeval with much of the plutonism in the South Lake Igneous Complex. The regional partitioning of hornblende diorite and quartz tonalite bodies is strikingly evident in the north of the igneous complex, where shear zones in diorite and tonalite plutons wrap around metagabbro enclaves. However, in the southern parts of the complex where metagabbro is absent, the relative disposition and local outcrop patterns of diorite and tonalite bodies are much more irregular and the mutual relations of such rock types are less readily discernible. The internal system of Ordovician shear zones is transected obliquely by the Siluro-Devonian faults which define the complex's external boundaries.

Emplacement of mesozonal arc tonalites and diorites was controlled by the earliest of the ductile fault structures preserved in the metagabbroic part of the South Lake Igneous Complex. Shear zones preferentially nucleated in the northernmost plutons were sited on preexisting shear zones within these ophiolite enclaves. Episodes of synplutonic tectonism are bracketed by the ages of the metagabbro and later dioritic and tonalitic plutons.

The Cambro-Ordovician movement history during ophiolite and arc development is revealed, in part, by the complex's mafic dyke rocks. Dyke swarms were sequentially emplaced in transiently dilatant structures that evolved during the brittle-ductile deformation of the constituent plutonic rocks. Moreover, as magmatic-state and solid-state deformation proceeded, synplutonic shearing was continuously focussed into progressively younger intrusions. Shear zone deformation overlapped, with declining influence, the emplacement of the intrusions comprising the South Lake Igneous Complex.

Siluro-Devonian Structural Relationships

The D_1 fault zones that bound the South Lake Igneous Complex dip steeply, generally to the east or the west. On a regional scale, the igneous complex and the bounding D_1 faults are openly folded by northeast-trending F_2 folds (Figure 45). They are also offset, in a brittle fashion, by later northeast-trending transcurrent structures of possible Devono-Carboniferous age (*see* MacLachlan, 1998). These features correspond with the major fault and fold structures mapped in adjacent stratified rock units (O'Brien, 1992).

The Siluro-Devonian fault zone along the eastern margin of the South Lake Igneous Complex is an imbricated overthrust zone in which the original stratigraphic order of rock units is reversed. Thus, in places, the Cambro-Ordovician South Lake Igneous Complex structurally overlies the Middle Ordovician Pennys Brook Formation, the Middle to Late Ordovician Lawrence Harbour Formation and the Late Ordovician Point Leamington Formation (Figure 45). Two small tectonic horses, one in the northeast and the other in the southeast part of the overthrust zone, illustrate the full lithotectonic sequence of inverted Ordovician rocks.

The Pennys Brook Formation, the youngest unit of the Wild Bight Group which is juxtaposed against the South Lake Igneous Complex, contains graptolites indicative of an earliest Caradoc age in the tectonic panel that lies northeast of the complex (*Nemograptus gracilis* Zone; S. H. Williams, 1991). The oldest known Point Leamington unit so disposed belongs to the latest Caradoc (*Dicranograptus clingani* Zone; S. H. Williams *et al*., 1992b). The Lawrence Harbour Formation black shale of mainly Caradoc age is mostly tectonically excised along the eastern boundary of the South Lake Igneous Complex. However, tectonic slices of this formation are locally preserved between detached fragments of the Pennys Brook and Point Leamington formations and, in one locality, the constituent black shales are directly overplated by the plutonic complex (Figure 45). In the structurally adjacent stratified units, fold asymmetry and axial surface inclination, together with the dip direction of attendant slaty cleavage and the pitch of a related extension lineation, indicate that the structures bounding the South Lake Igneous Complex are commonly high-angle reverse and oblique-slip faults or, more rarely, steeply dipping strike-slip thrusts.

The regional D_1 thrust fault at the western contact of the South Lake Igneous Complex crosscuts the intrusive boundaries of several early Ordovician plutonic units and transgresses the internal stratigraphy of the Pennys Brook Formation of the Wild Bight Group. In the north, the west-dipping D_1 thrust fault along the eastern margin of the igneous complex is interpreted to be displaced by the east-dipping D_1 thrust fault along the western margin of the igneous complex (Figure 45). In the vicinity of this juncture, a narrow, tightly F_2-folded, D_1 thrust slice of Lawrence Harbour black shale separates a panel of upper Pennys Brook strata in the northeast from a panel of middle Pennys Brook strata in the southwest. This tectonic sequence of stratified rocks intervenes between two discrete horses of the South Lake Igneous Complex (Figure 45).

Based solely on its structural outcrop pattern and the opposing dips of its bounding faults, the South Lake Igneous Complex could be interpreted as occupying several klippe resting above the stratified rock units. On regional considerations, however, it is more likely that the horses of plutonic rocks constitute part of a positive flower structure.

If the detritus in earliest Ashgill conglomerate of the Point Leamington Formation on Martin Lake was derived from South Lake tonalite (*see* section on Regional Interpretations of the Badger Group, page 57), then it is probable that the South Lake Igneous Complex is parautochthonous and that the amount of stratigraphic separation and thrust displacement is minor. The grain of the ductile Siluro-Devonian structures, particularly the northwest-trending set of bounding faults, may be inherited from the preexisting architecture of the plutons in the South Lake Igneous Complex.

Tectonic Linkages between the South Lake Igneous Complex and the Wild Bight Group

Dean (1977) stated that certain members of the South Lake Igneous Complex intruded the Wild Bight Group and, although this is not possible for adjacent Pennys Brook strata, it is a likely relationship for volcanic rocks found in the lowest exposed part of the group. In support of this contention, early Ordovician tonalite plutons are observed to contain silicified finely-layered and banded xenoliths that could be argued to belong, on lithological grounds, to the Wild Bight Group. However, these accidental xenoliths possess pre-incorporation structural fabrics (Plate 40). Thus, the conformity of the *N. gracilis* Zone and older Wild Bight volcanosedimentary sequence is difficult to rationalize with the implied early Ordovician ductile deformation of parts of the local Wild Bight Group succession.

Certain mafic dyke swarms in the arc-related mesozonal plutons of the South Lake Igneous Complex may potentially be correlated or tectonically linked to minor Ordovician mafic intrusions in the lower Wild Bight Group (MacLachlan, 1998). One group of gabbro, diorite and diabase intrusions which is possibly related to one of the late South Lake dyke suites occurs in swarms that are emplaced into the bimodal volcanic rocks of the lower Wild Bight Group. The Wild Bight host rocks are situated below lower Pennys Brook and older volcaniclastic strata, which contain conglomerates having conspicuous highly rounded granitoid clasts. These minor mafic intrusions were intruded after the silicic alteration of the mafic volcanic rocks in the lower Wild Bight Group (e.g., Figure 14).

The early Ordovician fabrics in the synplutonic shear zones of the South Lake Igneous Complex are not expressed in the geochemically similar and age-equivalent (MacLachlan and Dunning, 1998a) supracrustal rocks of the Wild Bight Group. This implies that these units did not reside in the same part of the crust in earliest Ordovician time, or that earliest Ordovician deformation was confined to certain magmatic conduits and was not regionally developed, or that only the youngest undeformed quartz-feldspar porphyry phases of the tonalite tracts were emplaced high enough to intrude the lower Wild Bight Group supracrustal rocks.

Summary of Ordovician and Silurian Relationships

The South Lake Igneous Complex records the mid-crustal evolution of a magmatic arc emplaced into a relict arc ophiolite of the eastern Dunnage Zone (MacLachlan and Dunning, 1998a). Several Ordovician periods of ductile deformation, dynamothermal metamorphism and mesozonal plutonism were focussed in precursor and contemporaneous shear zones during the rise of early Ordovician magmas.

Structurally, the South Lake Igneous Complex is a parautochthonous unit which is externally bounded by high-angle reverse faults of probable late Silurian age. These structures obscure original relationships with various stratified rock units of Middle to Late Ordovician age. Tectonic links that the Cambro-Ordovician South Lake Igneous Complex have with the Early to Middle Ordovician Wild Bight Group and Late Ordovician Point Leamington Formation are interpreted to mean that the South Lake Igneous Complex was proximal to the lower Wild Bight Group by the late Arenig and Llanvirn and that the South Lake Igneous Complex and the upper Wild Bight Group were proximal to the Point Leamington Formation by the late Caradoc.

POSTTECTONIC MAJOR INTRUSIONS

In the area surveyed, discordant masses of dominantly medium- to coarse-grained plutonic rocks comprise parts of three small Paleozoic batholiths, each of which is greater than 100 km^2 in area. Some of these batholiths, or parts of them, may be more appropriately called intrusive suites, igneous complexes or injection complexes. The western margin of the Loon Bay batholith occurs on islands in the Bay of Exploits within the eastern part of NTS map area 2E/6, the northern margin of the Mount Peyton batholith is situated near Norris Arm in the southern part of NTS map area 2E/3, and the eastern margin of the Hodges Hill batholith is located near Northern Arm in the western parts of NTS map areas 2E/3 and 2E/6. Their relative disposition and regional structural setting are illustrated in plan view in Figure 46. The inferred geometry of selected Loon Bay and Mount Peyton plutons is also drawn schematically in cross section in Figure 46, where relationships with early and middle Paleozoic country rocks are generally depicted.

The discordant Budgells Harbour stock is mainly composed of alkaline gabbro and pyroxenite of Late Jurassic–Early Cretaceous age (Helwig *et al*., 1974; Strong and Harris, 1974). An associated radial swarm of lamprophyre dykes contain granulite facies xenoliths which are thought to represent the deep basement of the Exploits Subzone (Fryer *et al*., 1997). Posttectonic Mesozoic plutonic rocks are widespread in the area surveyed but they are not discussed further in this report.

Certain plutonic bodies within the Loon Bay, Mount Peyton and Hodges Hill batholiths have observable intrusive contacts and they have either Silurian or Devonian absolute ages (Table 1; Figure 4). Other undated plutons are seen to intrude fossil-bearing country rocks as young as the Early Silurian or exhibit textures similar to adjacent terrestrial volcanic rocks of presumed Middle Silurian age. However, a few of the constituent plutons are known to be older than the depositional age of the youngest host rocks adjacent to these batholiths and are thus isolated from the country rocks that they originally intruded (e.g., the pre-Late Silurian foliated, mid Ordovician granite in the Hodges Hill batholith; Dickson, 1999; Table 1). Other plutons, partly fault-bounded and partly intrusive, are equivalent in age to some of the batholith's stratified country rocks but are older than the age of regional deformation and contact metamorphism of such rocks. Thus, these plutons could not have generated the hornfels observed adjacent to the batholith in which they occur (e.g., the latest Ordovician–earliest Silurian granite in the Mount Peyton batholith; Dickson, 1993; Table 1).

Loon Bay Batholith

Granodiorite, tonalite and granite plutons comprise most of the posttectonic intrusive rocks in the Loon Bay batholith. Some of these intrusions have been isotopically dated in the northeastern part of the unit, where they have yielded Late Silurian ages (Elliot *et al*., 1991). On islands in the Bay of Exploits, the main rock type is a massive, equigranular, hornblende–biotite granodiorite, which is flanked, to the south and east, by feldspar–megacrystic

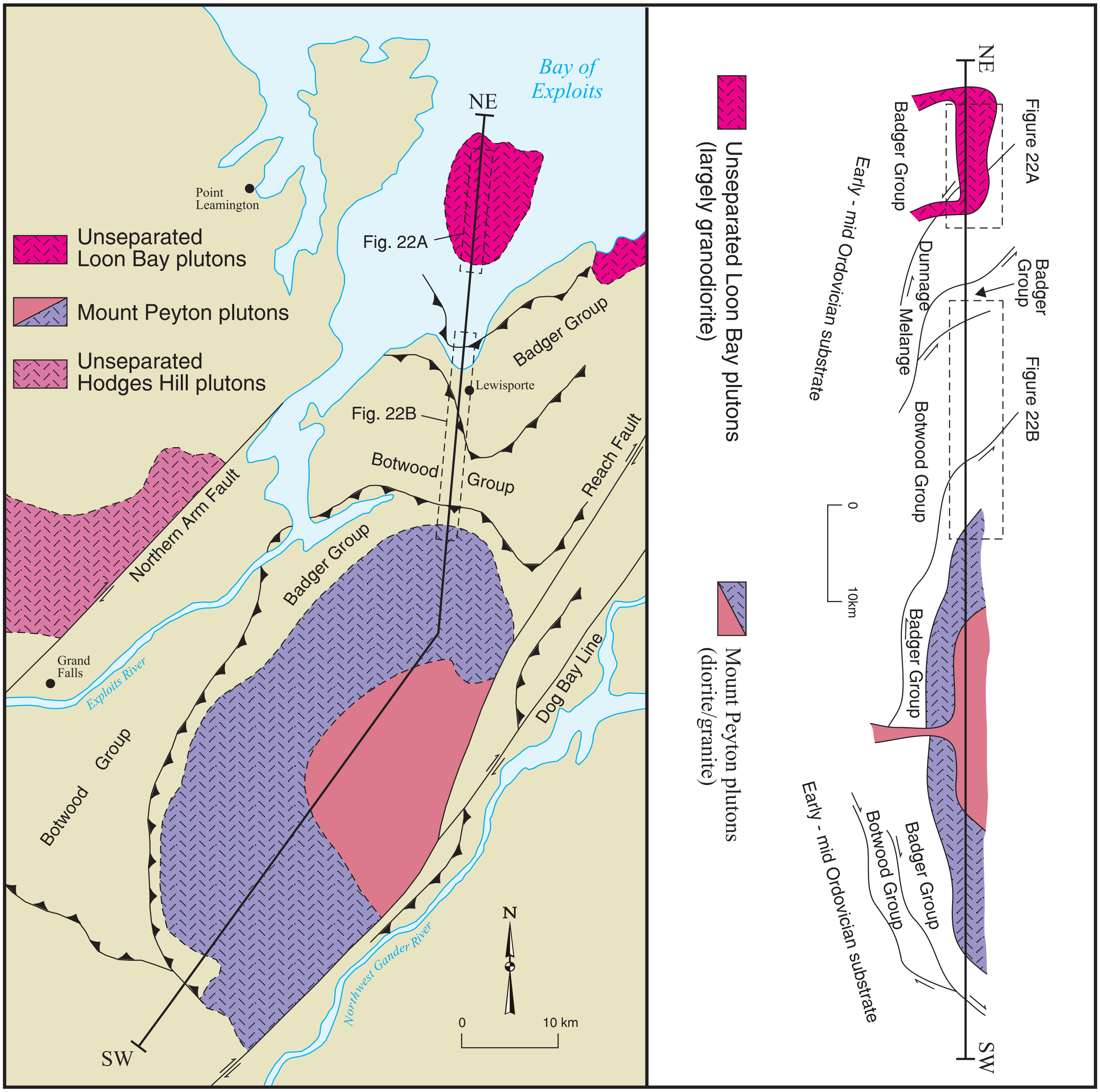


Figure 46. *Sketch map illustrating the regional geological setting and relative disposition of parts of the Loon Bay, Hodges Hill and Mount Peyton batholiths. Inset is a northeast–southwest cross section depicting the interpreted shape of the Loon Bay and Mount Peyton plutons and their relationships to regional structures in adjacent stratified rock units. The locations of Figures 22a and 22b are also illustrated.*

biotite granite. The megacrystic granite is schistose, in places, and is crosscut by quartz-feldspar porphyry and felsite dykes. It is also intruded by an unfoliated granodiorite that appears to be in transitional contact with the main hornblende - biotite granodiorite underlying Long Island and the adjacent archipelago (Figures 3 and 22a).

In the Indian Arm–Loon Bay area, Currie and Williams (1995) separated mappable bodies of quartz megacrystic, biotite K-feldspar granite (the Loon Bay Granite of Kerr *et al.*, 1991) and hornblende-biotite granodiorite gradational to tonalite and diorite. In contrast with the peripheral granitic plutons of the western Bay of Exploits, granodiorites occur at the margin of the batholith in the type area of Loon Bay. There, they are locally foliated and reported to intrude diorites and more mafic tonalites found in the central part of the batholith.

In western and eastern parts of the Bay of Exploits, moderately dipping intrusive sheets of granodiorite and granite trend northward and were injected along the batholith's western and eastern margin. Emplacement of these intrusions was probably contemporaneous with the east-trending biotite granite sheets observed at the southern margin of the Loon Bay batholith (Figure 22a), as indicated by mutual abutment relationships between intrusive sheets.

The granodiorite centred on Long Island in the westernmost part of the Loon Bay batholith has a crudely polygonal outcrop pattern (Figure 46; O'Brien, 1990). It displays a well-exposed, sheeted, semi-concordant boundary on the north and the south. However, the trace of its western and eastern margins are notably discordant to structures in the country rocks, even though such contacts are poorly exposed. Although the floor of this pluton has not been observed, its unique shape is thought to reflect the domed top of a horseshoe-shaped intrusive sheet (Figure 22a). It is possible that the Long Island granodiorite may also be horseshoe-shaped in a profile orthogonal to the one shown in this illustration, although the presence of the north-trending intrusive sheets complicates such reconstructions.

Field Relationships

Posttectonic dykes of quartz-feldspar porphyry are widespread along the north and south margins of the Long Island granodiorite, although most dykes occur outside of the pluton margin. Emplaced discordantly and concordantly into a variety of complexly deformed host rocks, the quartz-feldspar porphyries are observed to be strongly jointed and locally altered. At the northwestern contact of the Loon Bay batholith between Swan and Hornet islands, satellite porphyry dykes intrude highly sheared belts of tectonic melange developed within variably strained parts of the Cottrells Cove Group. Elsewhere in the Red Indian Line structural zone, the Long Island granodiorite is seen to intrude lenticles of Badger Group conglomerate, full of lithologically similar Ordovician or older granodiorite clasts.

At the southwestern contact of the Loon Bay batholith, in the Burnt Bay and Campbellton areas, undeformed Long Island granodiorite is present within the regional fault zone that separates the Dunnage Melange and the Badger Group (Figure 22a). Small sheets of biotite K-feldspar granite, some pristine and equigranular but others altered and foliated, are also emplaced into this fault zone. Near the base of the Dunnage Melange overthrust sheet on southern Upper Black Island, intrusive bodies of fresh gabbro cut microgranite sheets with narrow margin parallel mylonite zones. Within the South Hummock Island and North Upper Black Island thrust panels (Figure 22a), folded gabbro sheets and foliated porphyritic diabases intrude Upper Ordovician sedimentary rocks of the Badger Group. In some rare exposures, undeformed porphyritic diabase dykes are seen to intrude narrow shear zones in foliated Middle Ordovician gabbro sills emplaced into the volcanic rocks of the Lawrence Head Formation and other units of the Exploits Group.

Orthogneiss and Hornfelsic Schist

The oldest plutonic rock in the Loon Bay batholith in the Bay of Exploits area is a finely banded felsic orthogneiss. It forms a narrow belt along part of the southwest margin of the Loon Bay batholith, and is partially engulfed by the Long Island granodiorite. Either locally developed or incompletely preserved, orthogneiss is found along part of the northern thrust-faulted margin of the Dunnage Melange (Plate 41; Figure 22a).

Tectonic banding in the orthogneiss is truncated by megacrystic granite, and the intrusive contact between orthogneiss and megacrystic granite is cut by quartz-feldspar porphyry and late-stage microporphyritic granite (Plates 42 and 43). The absolute age of magmatic zircons from the oldest and youngest of these intrusions (G. R. Dunning, unpublished data, 1994) indicates that all these granitoids crystallized in the Early Devonian within a narrow time interval around the Lochkovian–Pragian boundary (cf., Tucker, R.D. and McKerrow, W.S., 1995; Tucker *et al*., 1998).

Coeval with or younger than the orthogneiss are concordant sheets of porphyritic granite which, in low-strain pods, are observed to be complexly infolded with wacke, limestone, basalt and pebbly mudstone, all metamorphosed to amphibolite grade. Near high-strain zones, these porphyritic granites are mylonitized, especially at the margins of the intrusive sheets. Calcsilicate-bearing psammite, aluminosilicate-bearing sulphidic pelite and garnetiferous amphibolite–the hornfelsic schist envelope surrounding the narrow belt of orthogneiss (Figure 22a)–are also observed to be tectonically straightened. The folded zones and the mylonite tracts are both crosscut by equigranular granitic intrusions. In places, several crosscutting phases of porphyritic and equigranular granitoids are seen to be strongly to weakly foliated near the steeply south-dipping mylonite zones developed in the contact aureole.

Where the megacrystic granite and equigranular granodiorite phases of the southern Loon Bay batholith have stoped contacts, thermal metamorphism of local country rocks or xenoliths is not evident. In other locations, a static metamorphic aureole between 100 to 200 m wide is present, especially where the host rock contact is sharp and the pluton is undeformed. In several places, the contact zone is up to 400 m in width and consists of alternating sediment screens and granitoid sheets, each no more than 20 m wide. There, spotted country rocks display raised podiform nodules and resistant calc-silicate ribs. It is possible that these hornfelsed screens were deformed by map-scale folds; alter-

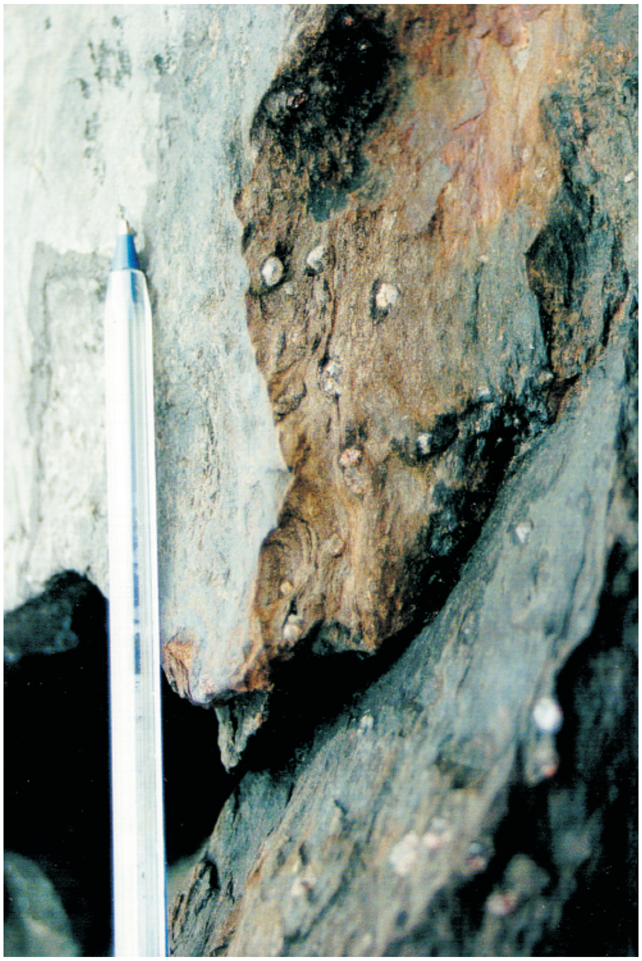


Plate 41. *Plan view of steeply dipping schistosity and augened garnet porphyroblasts in coarse-grained platey hornfels adjacent to the orthogneiss and gneissic granite of the Loon Bay batholith. The illustrated hornfelsic schists were developed in the matrix pelite of the Dunnage Melange on Birchy Island near Shoal Tickle.*

natively, granite sheets might have been guided by the openly folded shape of the host sedimentary strata. However, in marked contrast, the syntectonic contact aureole of the orthogneiss contains spotted cordierites with large aspect ratios, sheared pseudomorphs after andalusite crystals and augened garnet porphyroblasts (Plate 41).

Hodges Hill Batholith

The dioritic and granitic rocks of the Hodges Hill batholith are thought to have been derived from melts originating in the mid-Paleozoic mantle or within a juvenile magmatic arc (Fryer *et al*, 1992). Although they were emplaced into the composite oceanic-continental crust of the Dunnage Zone, they were not significantly contaminated by the sialic crust which, by the Late Silurian, had completely underplated the regionally extensive tracts of simatic crust in central Notre Dame Bay. In this regard, the Hodges Hill intrusions are isotopically contrasted with other similar aged plutonic rocks in the Gander Zone and also with other Exploits Subzone plutons located farther east (Fryer *et al*, 1992).

The Hodges Hill granite and the Twin Lakes diorite complex, informal map units originally surveyed by Hayes (1951), are included in the Hodges Hill batholith by most modern workers. In addition to these subjacent plutons, there are younger (e.g., Unit 4 of Kerr, 1995) and older (e.g., Unit 12 of Dickson, 1998) granitoid rocks present in the batholith. Hayes (1951) mapped amphibole-rich gneiss, hornblendite and pyroxenite and included them within the western part of his diorite complex. Mafic gneiss and amphibolites have not, however, been recognized farther east in the area surveyed.

The Hodges Hill batholith contains a variety of mafic and felsic plutonic rocks in the New Bay Pond–Northern Arm area, the region northwest of the Northern Arm Fault in Figure 46. With the exception of a tract of ground in NTS map area 2E/4 surveyed by Dickson (1998, 1999), they have not been separated on geological maps of the region. Some granitoid rocks in this part of the batholith are observed to be older than the ubiquitous diorite intrusions, whereas others are demonstrably younger. In the area surveyed, the Northern Arm pluton is reported to contain quartz monzonite, granodiorite and gabbro (Dean, 1977). In this regard, the plutonic body provides a small scale example of the heterogeneity of the Hodges Hill batholith. The Northern Arm pluton, which is locally cataclastically deformed, is terminated in the southeast by the Northern Arm Fault (Figure 46).

In the Northern Arm area, equigranular medium-grained diorite and subordinate fine-grained gabbro are intruded by hornblende–biotite granodiorite, itself transitional with more mafic intrusive rocks. In most places, the intrusions of gabbro, diorite and granodiorite appear to be undeformed, at least on the scale of the exposures examined. The gabbro–diorite bodies display chilled margins where they intrude equigranular, medium-grained, biotite granite and pink vuggy granophyre. Porphyritic microgranite and granophyre are also observed in direct contact with volcanic country rocks (Plate 44). In contrast, the granodiorite commonly contains variably digested xenoliths of much coarser granitoid rocks.

Field Relationships

Abundant swarms of plagioclase porphyritic diabase dykes (Plate 45) are present where fresh gabbro intrudes epizonal granite; however, they are also widespread in Ordovician and Silurian country rocks outside the Hodges Hill batholith. The former relationship distinguishes the unfoliated, granite-hosted mafic dyke swarms (Figure 17) from the folded Ordovician diabase dykes and associated gabbro-diorite sills in the Exploits and Wild Bight groups (*see also* Dickson, 1998). Immediately northeast of New Bay Pond, vertical dykes of plagioclase porphyritic diabase trend northeastward across several, steeply dipping, northwest-trending rock units belonging to the Exploits Subzone (Figure 17). Widely developed in and near the Hodges Hill batholith, Hayes (1951) mapped similar diabases and dyke relationships west of North Twin Lake within host rocks now assigned to the Notre Dame Subzone or the Red Indian Line structural zone.

Sheeted gabbros with the same orientation as the plagioclase porphyritic diabases are seen to intrude thermally metamorphosed Late Ordovician sedimentary strata of the Gull Island Formation of the Badger Group. They also crosscut early Ordovician volcanic strata in the adjacent New Bay Pond sequence of the Wild Bight Group. These gabbro sheets are assigned to the Hodges Hill batholith along with the plagioclase porphyritic diabase intrusions.

In the New Bay Pond area, fresh equigranular microgranite is seen to intrude medium-grained ophitic gabbro. Both rock types are undated but previous workers have grouped these intrusions with the Hodges Hill intrusive complex or its satellite bodies (Kusky, 1985). They are generally similar to the felsic and mafic plutonic rocks located immediately north of the main batholith between Rowsells Lake and New Bay River, where they comprise small stocks and bosses hosted by rocks of the Wild Bight Group (e.g., Unit 22 of O'Brien, 1993). The older gabbroic bodies are unfoliated and, south of New Bay River, they have xenoliths of silicified pyritic basalts, as do certain Ordovician intrusions. However, it is the presence, east of New Bay Pond, of hornfelsed metasedimentary rock enclaves that probably indicates a Siluro-Devonian age for such gabbros.

Plate 42. *Plan view of crosscutting, steeply dipping intrusive contact between unfoliated feldspar-phyric biotite granite and well-foliated granodioritic orthogneiss at the margin of the Loon Bay batholith on the southwest coast of Birchy Island.*

Plate 43. *A highly discordant body of quartz-feldspar porphyry, similar to the satellite intrusions of the Loon Bay batholith, crosscuts the slightly discordant intrusive contact of biotite granite with banded orthogneiss on the southwest coast of Birchy Island.*

Within the easternmost part of the Hodges Hill batholith, the younger granites are gradational with mafic–felsic intrusive breccias, and they are host to the plagioclase–porphyritic diabase dyke swarms. The aplite-veined pink microgranites are crosscut by steeply dipping tuffisite pipes, which contain xenoliths of pink ignimbrite (Plate 44). The microgranites are also spatially associated with subvertical quartz-feldspar porphyry bodies that display strong flow-layering parallel to their intrusive margin. Similar flow-banded felsic dykes occur south of the Loon Bay batholith near Powderhouse Cove and Masons Cove, where they are reported to be gold-bearing (Figure 18).

Plate 44. *Plan view of crosscutting intrusive contact of feldspar-porphyritic microgranite (below lens cap) with steeply dipping, flow-banded rhyolite at the margin of the Hodges Hill batholith near New Bay Pond. Country rocks are Charles Lake volcanics of the Silurian Botwood Group.*

Plate 45. *Plan view of the chilled margin of a vertical plagioclase-porphyritic diabase dyke which intrudes a fine-grained granite porphyry unit within the Hodges Hill batholith near the Northern Arm Fault. Along an irregular step-shaped contact, the diabase is seen to have occupied right-hand en echelon microfractures in country rock granite.*

Late-stage Faulting of the Batholith

An unfoliated equigranular granodiorite of the Hodges Hill batholith displays decimetre-wide zones of cohesive fault breccia near the Devono-Carboniferous Northern Arm Fault (Figure 17). However, such breccia is also found within the batholith's Early Ordovician to Late Silurian country rocks adjacent to and in the general vicinity of this transcurrent fault structure. Rock units known to contain cemented breccia zones include the Northern Arm Basalt of the Wild Bight Group, a diorite pluton of the Phillips Head igneous complex, several of the lower chert units of the Shoal Arm Formation and some redbeds in the Wigwam Formation of the Botwood Group.

Most of the cohesive fault breccia zones developed near small dextral strike-slip shears (Kusky *et al*., 1987) that dip vertically and trend northeastward (Figure 17). They are typically located near brittle rectilinear fractures which are readily observable on aerial photographs. However, some breccia zones are seen in association with a conjugate set of northwest-trending sinistral shears. Gently northwest-dipping breccia zones hosted by Hodges Hill granodiorite and Botwood red-beds may relate to less common, oblique-slip thrust faults. These possibly indicate the position of mineralized positive flower structures along the Northern Arm Fault (Figure 17).

Between New Bay Pond and Northern Arm, several breccia zones illustrate comminuted blocks and fine-grained matrix that are both highly silicified. Some are monolithic cataclastic breccias; whereas, others may be heterolithic explosion breccias. Fault lineaments associated with such breccias are also the preferred site of intrusion of the regional swarm of plagioclase-porphyritic diabase dykes. Some diabase dykes in the Hodges Hill swarm are altered and intruded by quartz vein arrays (Figure 17); others are fresh and crosscut the silicified breccia zones.

Mount Peyton Batholith

The northern margin of the Mount Peyton batholith is dominantly composed of posttectonic rocks of gabbroic and granitic composition. The predominant rocks in the northern part of the Mount Peyton intrusive suite (Blackwood, 1982; Dickson, 1992) are homogeneous hornblende-bearing diorite and massive pyroxene-rich gabbro. Granite, granodiorite and tonalite are less common, although they comprise mappable units in certain areas of the batholith (Dickson, 1992). Strong and Dupay (1982) suggested different origins for the gabbros and the granites in the bimodal Mount Peyton suite. They concluded that mantle-derived older gabbros melted the lower crust to provide melts which fed the younger granites. Continuous fractional crystallization of a calc-alkalic intrusive suite was considered unlikely.

At Rattling Brook, near the northern boundary of the batholith, an isotropic grey diorite has been reported to have yielded abundant late Silurian magmatic zircons (G. R. Dunning, unpublished data, 1994). As gently dipping folded beds extend some 3 km from the margin of this intrusion to a late Llandovery fossil locality (Boyce and Ash, 1994), the intruded strata are presumed to include some of the youngest known early Silurian rocks in the Badger Group (Figure 22b). The Rattling Brook diorite is similar in age to the gently dipping, layered sheets of gabbro, diorite and troctolite (Eckstrand and Cogulu, 1992, 1993) which outcrop near Rolling Brook and throughout the southern part of the batholith (Dickson, 1992).

Some granitic bodies included in the Mount Peyton batholith have yielded isotopic ages older than the Rattling Brook and Rolling Brook diorites and they are thus distinguished from the post-diorite granitic bodies found near the batholith's northern margin (Dickson, 1994; Table 1). Located farther southeast in the batholith near the extension of the Reach Fault and the Dog Bay Line, such late Ordovician–early Silurian granitoids are faulted against a variety of stratified rock units. They have temporal counterparts in the older parts of the Topsails intrusive complex in west-central Newfoundland (Whalen, 1989) and plutons of similar age are also present in the New England and New Brunswick Dunnage Zone (Tucker and Robinson, 1990; van Staal, 1994).

The Southwest Brook pluton is assumed to be a satellite body of the Mount Peyton batholith. It is composed of a fine-grained biotite granite which is transitional to porphyritic microgranite and granophyre. All these rocks are seen to be intruded by a suite of aplite dykes which are also present at the margin of the main batholith. Approximately 10 km long and 2 km wide, the Southwest Brook granite is seen to intrude Badger Group strata as young as Hirnatian in age (Boyce and Ash, 1994). The pluton is generally undeformed, although the granite is tectonically brecciated near the fault contact between the Badger and Botwood groups.

Structural Setting of the Northern Plutons

In the area surveyed, the mafic and felsic plutonic bodies of the Mount Peyton batholith are known to have crosscutting intrusive relationships with previously folded and ductilely thrusted sequences of Ordovician and Silurian turbidites. However, none of these intrusions have been seen in direct contact with the younger tract of Botwood Group redbeds and associated volcanic rocks that partially surrounds the Mount Peyton batholith (Figure 46). Nevertheless, since the mutual boundaries of the terrestrial Botwood Group and the marine Badger Group are regional D_1 thrusts and, as minor D_1 structures related to these faults are locally transgressed by gabbro, both groups of Silurian strata are assumed to be intruded by the posttectonic pluton (Figure 22b). Such gently dipping fault structures are postulated to have guided mafic pluton emplacement at depth (Figure 46 cross-section).

The batholith's mafic and felsic plutons, satellite bodies and hornfelsed country rocks are reported to be deformed by brittle transcurrent faults near the Northwest Gander River (Tallman and Evans, 1994; Figure 46). However, this displacement postdated the climactic tectonometamorphic events recorded in east-central Notre Dame Bay (e.g., the Reach Fault of Currie and Williams, 1993; Williams *et al*., 1993). In the south-central part of Notre Dame Bay, such brittle structures are assumed, though not observed, to have affected the Botwood Group as well as the underlying rock units.

In contrast with the interpretation shown in Figure 22b, it is possible that some southwest-dipping brittle thrusts offset diorite and also displace the Botwood Group. This would make the Mount Peyton diorite pluton and its country rocks structurally akin to the plutonic and stratified rocks comprising the flower structure near the transcurrent Northern Arm Fault (Figure 17). However, it would also imply that brittle thrusts in the posttectonic diorite are essentially coplanar with similarly oriented ductile thrusts in the Badger Group. In this explanation, the ductile thrusts in the Badger Group could predate the deposition of the Botwood Group; whereas, the brittle thrusts could represent the initial structures to deform the Wigwam Formation and the Mount Peyton diorite.

Field Relationships

Dark grey to green, equigranular, unfoliated diorite is systematically jointed and well fractured. On aerial photographs, lineaments and major fractures are observed to trend dominantly northeastward and northwestward. Near certain joints, primary plagioclase is altered to saussurite and primary hornblende and pyroxene are replaced by chlorite. Elsewhere, coarse-grained hornblende-bearing pegmatites of presumed magmatic origin develop along joints within fine- to medium-grained diorite.

The diorite forming the northernmost part of the Mount Peyton batholith is intruded by small, fine-grained bodies of

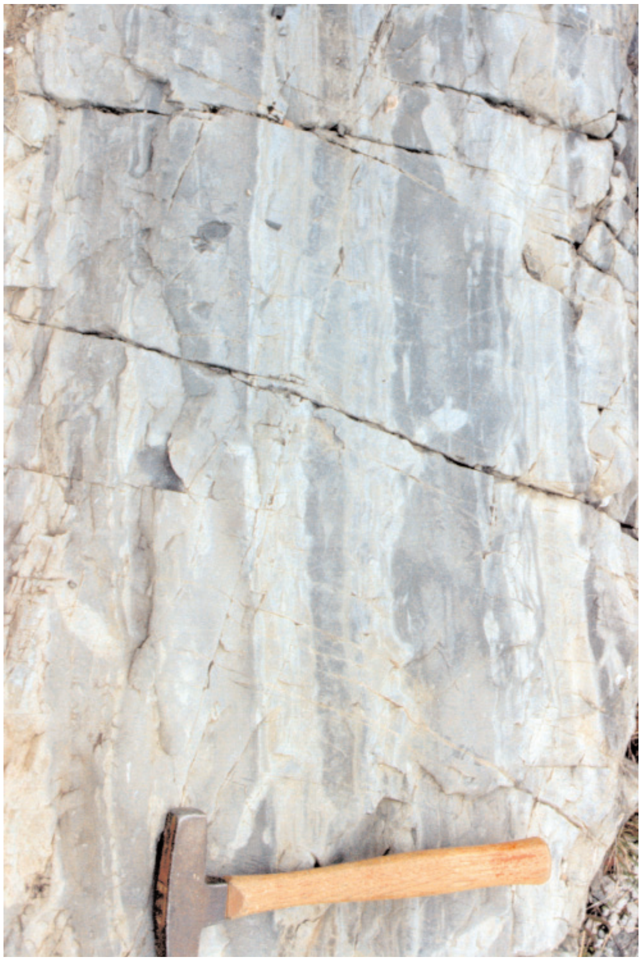


Plate 46. *Primary bedforms of Badger Group turbidites are preserved in a nodular spotted hornels within the outer thermal aureole of the Mount Peyton diorite near the mouth of Rattling Brook.*

Plate 47. *Agmatite zone east of Rattling Brook within the inner thermal aureole of the Mount Peyton diorite is marked by randomly oriented, variably digested blocks of metasedimentary restite (below hammer). A relatively cohesive granofels zone in the Badger Group contains porphyroblast-rich, relict-bedded nodular rocks (above hammer) that are seen to have flowed and been variably rotated adjacent to the agmatite zone.*

biotite granite, in places, with mariolitic cavities and associated pegmatite. Such leucocratic granites form small isolated plug-like intrusions within the strongly jointed and microfaulted diorite or else they comprise steeply dipping, northeast-trending, lineament-parallel sheet intrusions (Dickson, 1994; Dickson *et al*., 1995).

Satellite bodies of muscovite- and biotite-bearing granite are found in metasedimentary country rocks adjacent to the diorite. These pegmatitic granites are most common within granoblastically recrystallized or agmatized parts of the marginal coarse-grained hornfels.

Metamorphic Aureole of the Mount Peyton Diorite

A spectacular zone of granofelsic migmatite, up to 2 km wide, occurs within the thermal aureole of the Mount Peyton diorite in certain areas (e.g., Norris Arm). In such localities, the very outer part of the aureole includes a dark-coloured, weakly spotted or baked argillaceous hornfels (Plate 46). Closer to the pluton, fine-grained metasedimentary rocks have the regional slaty cleavage preserved as fine inclusions in porphyroblasts but otherwise display a granofelsic texture. Such metaturbidites pass laterally and up-section into medium-grained granoblastic migmatite and hornfelsic schist. In the extensively recrystallized, highest grade, innermost part of the contact metamorphic aureole, relict sedimentary layers display a completely isotropic matrix devoid of internal bedforms. Such rocks are transitional to zones of coarse-grained agmatite (Plate 47).

Although the agmatite matrix may be represented by fine-grained granodiorite in places (Dickson, 1994), the unfoliated matrix of the granofels is commonly *in situ* sedimentary rock which has been mimectically recrystallized.

The agmatitic portion of such quartzofeldspathic metamorphic rocks has been statically grain-coarsened at the expense of the foliated restite.

In numerous localities, the matrix of the migmatite is seen to preserve thin layers of Bouma turbidite divisions marked by podiform concretions and discontinuous calc-silicates, each several centimetres in diameter. However, in other places, medium-bedded strata are observed to comprise blocks up to 10 m by 15 m in size and form spectacular partially disaggregated trains of country rocks within the migmatite zone. Individual metasedimentary blocks are body rotated relative to each other and the regional tectonic grain. Nevertheless, a ghost turbidite stratigraphy is locally present within the aureole of the Mount Peyton diorite.

The granites in the northern part of the Mount Peyton batholith were apparently intruded at a relatively late stage in the development of the thermal aureole. Highly metamorphosed rocks from the inner aureole occur as discrete, randomly oriented xenoliths within biotite–muscovite granite. Thus, these granite bodies crosscut the diorite and its underlying aureole (Figure 22b). In contrast, where the aureole is narrow, a pre-granite phase of hornblende diorite contains both sharp- and diffuse-edged enclaves of spotted hornfels having randomly oriented amphibole porphyroblasts. It is probable that the diorite rose to stope its own aureole and, therefore, metamorphic enclaves in diorite may represent the protolith of the granofels and agmatite zones developed elsewhere.

Geophysical modelling of the Mount Peyton batholith suggests inward-dipping blocks of mafic plutonic rocks overlain by a relatively thin veneer of felsic plutonic rocks (Miller and Thakwalakwa, 1992). This configuration is adopted in the cross section in Figure 46. Their gravity profile indicates that the basal mafic intrusive rocks display their most gentle regional dip along the northwest margin of the complex, where the bottom of the batholith is inclined southeastward from the surface to depths of about 3 km. This is consistent with the diorite–gabbro bodies occupying or crosscutting gently dipping thrust sheets in Badger Group turbidites (Figure 46), or with part or all of the marginal phases of the batholith being themselves reverse faulted near a post-thrust system of brittle transcurrent faults.

In places, the intrusive or hornfelsed rocks causing the magnetic high over surface exposures of the northwest contact of the batholith have a subvertical dip rather than a gentle southeastward dip. To explain this phenomenon, Miller and Thakwalakwa (1992) suggested that the gabbro sheet intrusions were offset by later northeast-trending upright structures, thereby giving them sharp edges in cross section. Moreover, these authors postulated that Silurian-aged gabbros were affected by a period of brittle faulting immediately prior to the intrusion of Devonian aged granites. This Acadian tectonic disturbance was argued to possibly also account for intrabatholith discrepancies in the remnant magnetization, magnetic susceptibility and paleomagnetic poles as well as the resetting of the Ar^{39}/Ar^{40} isotopes in gabbro and diorite.

Dickson (1994) interpreted the contact of the Mount Peyton gabbro to dip gently northwestward beneath the openly folded turbidites of the Badger Group near Norris Arm. Closer to Rattling Brook, O'Brien (1993a) regarded the contact of the Mount Peyton diorite to dip gently southeastward above the hornfelsed turbidites of the underlying Badger Group (Figure 46). As Mount Peyton gabbros and diorites were probably emplaced in gentle sheets near various thrust slices or open folds, the ponding of magma and the slow transfer of heat may have locally caused high-grade contact metamorphism along parts of the intrusive contact.

Based on Dickson's (1992) and Dickson and Colman-Sadd's (1993) mapping, most of the granite plugs are elongated northeastward. They appear to have been emplaced along brittle conduits parallel to the Figure 46 section plane more commonly than they were along conduits perpendicular to this section plane. It may be significant that not all of the marginal granite sheets and pegmatites cut through the country rock migmatite and adjacent massive diorite with vertical dips. Intrusive granite sheets inclined gently northwestward and southeastward are also seen to separate the main diorite body from the aureole granofels.

IMPLICATIONS FOR MINERAL EXPLORATION IN CENTRAL NOTRE DAME BAY

The known massive sulphide occurrences in Notre Dame Bay are associated with lithostratigraphic units of different age, lithodemic association, subzone affinity and tectonic environment (Swinden, 1990, 1991). As massive sulphide deposits occur to the south (Walker and Collins, 1988) in probable Early Ordovician strata of the Exploits Subzone and to the west (Thurlow, 1996) in probable Middle Ordovician strata of the Notre Dame Subzone, correlative or lithologically similar units in the study area are highly prospective for volcanogenic base metals. Targets traditionally explored have been the boundaries between basalt and less voluminous rhyolite horizons in marine sequences of pericratonic island-arc rocks in the Notre Dame Subzone and primitive island-arc rocks in the Exploits Subzone.

Far less prospected are the late Arenig–early Llanvirn hemipelagic and volcaniclastic sedimentary and oceanic basalt successions that are underlain by or, in places, interstratified with calc-alkaline basalts. In the Exploits Subzone, these include the upper New Bay Formation, Lawrence Head Formation and Strong Island chert of the Exploits Group, the Loon Harbour volcanics and lower Baytona chert of the Campbellton sequence, the Pennys Brook Formation of the Wild Bight Group, the volcanic rocks of the Phillips Head igneous complex, the volcanic rocks and limestone lenses of the upper Summerford Formation, and the unbroken formations of the Boones Point–Sops Head melange complex (Table 1). Such units represent the lateral equivalents of continental-margin arc volcanic rocks to the east, west and farther south. In contrast, age-equivalent or lithologically similar sequences in the Notre Dame Subzone overlie and may interdigitate with volcanic rocks erupted within a mature continental margin arc setting (e.g., the oceanic basalts, limestones and volcaniclastic turbidites of the Moores Cove Formation of the Cottrells Cove Group and the Crescent Lake Formation of the Roberts Arm Group).

Some of these volcano-sedimentary successions are found above much older primitive arc basalts having massive or stockwork sulphides (e.g., the mid-upper Exploits Group formations above Tea Arm-type volcanic and intrusive rocks, or the mid-upper Wild Bight Group formations above Glovers Harbour-type volcanic and intrusive rocks). Others are inferred to overlie the gabbro and sheeted dyke horizons in some of the region's purported arc ophiolites (e.g., the mid-upper Wild Bight Group formations above the South Lake Igneous Complex, or the mid- upper Mortons Harbour Group formations above the Betts Cove (?) igneous complex). The altered basaltic, volcaniclastic and hemipelagic strata occur as overstep sequences above remnant arc volcanic rocks on *both* sides of the Red Indian Line, and they have significant potential for Fe, Mn, Cu and Pb mineralization (Swinden *et al*., 1997; MacLachlan, 1998).

The most prospective early to mid Ordovician strata occur adjacent to early Ordovician volcanic and intrusive centres located in the northwestern Cottrells Cove Group and the southeastern Wild Bight Group. The regionally extensive, oxygen-rich, low-temperature alteration of these sedimentary and volcanic rocks produced vein systems and porphyroblasts which began to form during the compaction history of the altered horizons (e.g., Rees, 1999). In places, silicified and mottled volcaniclastic strata of the middle Exploits Group, which are more distal turbidites than those seen in the Wild Bight Group, also illustrate evidence of this sub-upper greenschist facies alteration (Rees, 1999). In the Exploits Subzone, overlying middle-late Ordovician pelagic black shales are pyrrhotite-bearing where highly reducing conditions prevailed in the depositional basin and, locally, such sulphides are known to be auriferous (Dean and Meyer, 1983).

Some ductile fault zones in central Notre Dame Bay are occupied by deformed sheets of gabbro and granite, and these mid-Paleozoic structures have potential for shear zone-related, iron-carbonate alteration and associated precious-metal mineralization (Evans, 1996). Other gold and antimony prospects are probably located along lineaments associated with brittle transcurrent faults (Tallman and Evans, 1994). Displacing posttectonic plutons and terrestrial country rocks, the antimony-bearing breccia veins adjacent to transcurrent structures are geologically younger features than the gold-bearing ductile shear zones.

Several different types of mineral deposits are potentially present in the map units surveyed in the central Notre Dame Bay region because a wide variety of rocks are observed to be altered. Multiple periods or styles of mineralization are inferred based mainly on the stratigraphic framework, geochemical signature or structural setting of the host unit.

UNITS WITH POTENTIAL FOR VOLCANOGENIC BASE-METAL DEPOSITS

Disseminated and vein-hosted chalcopyrite mineralization occurs locally within variably pyritized and silicified units of mafic volcanic rocks in the Moretons Harbour, Cottrells Cove, Exploits and Wild Bight groups. In many locations, where the original mineralogy of the mafic lava has been completely replaced and overgrown by matrix sulphides, the ghost outline of selvages of negligibly strained pillows are preserved. In all but the Moretons Harbour Group, thin units of laterally discontinuous, felsic pyroclastic rocks–some altered and some pristine–are spatially associated with underlying units of pillowed basalt, which typically host cupriferous or pyritic stringer mineralization. However, the prospective intervals of mafic and felsic rocks in the Exploits and Wild Bight groups have a distinctly different petrochemical signature than those in the Cottrells Cove Group (Swinden, 1988; O'Brien *et al*., 1997; Dec *et al*., 1997), despite the fact that the age and local geological setting is similar (Swinden, 1991; MacLachlan and Dunning, 1998a). Regionally, the alteration zones in each of the above rock groups follow the folded boundary between the mafic and felsic units, which suggests that the mineralization is essentially in the form of stratabound stockworks.

In the stratigraphically lower parts of the Cottrells Cove, Exploits and Wild Bight groups, altered pillow lavas are capped by relatively thin horizons of iron-rich chert, siliceous argillite and thixatropically deformed mudstone. These split laterally to separate relatively thick successions of proximally sourced volcanogenic turbidites that pinch out toward the volcanic centres. However, in all these groups, it is common for later thrust faults to structurally isolate the strata with massive sulphide potential and to juxtapose them against other sequences which are less prospective for volcanogenic base-metal mineralization.

Western Head Formation of the Moretons Harbour Group

Numerous pyrite, minor arsenopyrite and rare chalcopyrite showings are found in the Moretons Harbour Group on the Fortune Harbour Peninsula. Most of these mineral occurrences are located near the stratigraphic boundary between Sweeney Island Formation basalt and its sheeted dyke swarm and the overlying pillow breccias, tuffs, cherts and ferrugenous sediments of the Western Head Formation. Despite being generally stratabound in nature, Dean (1977) regarded this disseminated and vein-type hydrothermal mineralization as probably being somewhat younger than the age of the adjacent host rocks. In the type area of the Moretons Harbour Group, in a sequence composed totally of mid-ocean ridge or ocean-island tholeiite (Swinden, 1996), most epigenetic base metal and associated granophile element mineralization occurs at stratigraphic levels well below the Western Head Formation but still above the main sheeted dyke swarm in this rock group (Swinden *et al.*, 1988).

Fortune Harbour Formation of the Cottrells Cove Group

Iron, manganese and copper occurrences lie within or adjacent to the sedimentary subunits of the Fortune Harbour Formation of the Cottrells Cove Group. The stratigraphic succession is dominated by thin-bedded sequences of fine siliciclastic sandstone and mudstone interlayered with several types of primary and secondary chert. However, intervals of slumped debrites, isolated basalt flows and coarser-grained pyroclastic and epiclastic turbidites are also present (Dec *et al.*, 1997). The debrite matrix contains broken crystals of resorbed quartz and euhedral feldspar; whereas, the larger blocks and rafts are represented by chilled pieces of comminuted pillow lava, fractured beds of intraformational chert and partially-lithified felsic tuffs. Most of this detritus formed within the Fortune Harbour depositional basin and was not appreciably fragmented during transport in these volcanogenic mass-flow deposits. Sulphide clasts were not observed.

Iron–Copper Occurrences

The abandoned Grey Copper Mine and the Mitchell iron prospect (Dean , 1977; Figure 47) are hosted by mafic pillow breccia and occur along strike within culminating periclinal folds south of Fleury Bight. Mineralization occurs in folded and sheared veins that are enriched in chalcopyrite, pyrite and hematite relative to their host rocks. Basalts at both localities comprise part of a well-exposed belt of fault-bounded volcanic rocks outcropping between the villages of Fortune Harbour and Moores Cove. These mineralized lavas and pillow breccias have geochemical signatures indicative of an island arc origin (O'Brien *et al*., 1994; Dec and Swinden, 1994), although none show the LREE- and Eu-depletion characteristic of basalts associated with high-silica rhyolites and massive sulphide lenses (Swinden *et al.*, 1989; Galley, 1995). Interstratified sedimentary rocks typical of the Fortune Harbour Formation are notably absent from this netveined belt of altered mafic volcanic rocks.

Iron–Manganese Occurrences

Felsic pyroclastic rocks are most voluminous and coarsest grained in the northwestern part of the Fortune Harbour Peninsula. There, a felsic volcanic unit of the Fortune Harbour Formation stratigraphically overlies the basalts found at the Grey Copper Mine and the Mitchell iron prospect. Farther east, this subunit's felsic ash tuffs are interstratified with red, green and grey beds of chert, siliceous argillite and tuffaceous wacke. At least some Fortune Harbour felsic volcanism and sedimentation is Early Ordovician in age, as these tuffs have been isotopically dated at 484 ±2 Ma in one location (Dec *et al*., 1997).

Thin-bedded or laminated cherts are red (hematitic), dark grey (manganiferous) and turquoise (celadonitic or ferrophengitic). Some ribbon cherts in the Fortune Harbour Formation have preserved relics of recrystallized microscopic radiolaria, indicating the presence of some original biogenic chert (Dec *et al*., 1993; Dec and Swinden, 1994). However, other Fortune Harbour Mn- and Fe-rich hemipelagic cherts and siliceous pelagic mudstones may have formed as chemical precipitates near a Tremadoc-Arenig fluid discharge zone (Swinden, 1976). Certain size-graded silicified turbidites in the Fortune Harbour Formation originally accumulated by traction on the seafloor; however, they were replaced by variegated inorganic chert. Regardless of origin, these distal volcaniclastic rocks are host to the Sweaney's manganese prospect (Figure 47).

The felsic volcaniclastic sequence at the Sweaney's prospect can be traced along the flank of a sheared syncline underlying Southeast Arm to the abandoned Cookstown iron mine in the village of Fortune Harbour. Here, and at other localities along the trace of the Chanceport Fault, sporadic pyrite stringers are found in a variety of volcanic, sedimentary and mafic intrusive rocks.

Copper–Arsenic Occurrences

Chalcopyrite and arsenopyrite mineralization has been reported immediately north of the Chanceport Fault at Little North Harbour on the Fortune Harbour Peninsula and near Tom Wall Harbour on Exploits Island (Dean, 1977; Figure 47). The local host rocks have been variably assigned to the Western Head Formation of the Moretons Harbour Group (O'Brien, 1990) or the Fortune Harbour Formation of the Cottrells Cove Group (Dec *et al*., 1997). Both units are known to be mineralized and to contain very similar volcanic and sedimentary lithodemes.

The red- and grey-coloured beds of chert, ash tuff and siliceous argillite at Little North Harbour comprise relics of the stratigraphically highest rock unit preserved in the study area north of the Chanceport Fault. By comparison, on the northern side of this same fault structure in the Moretons

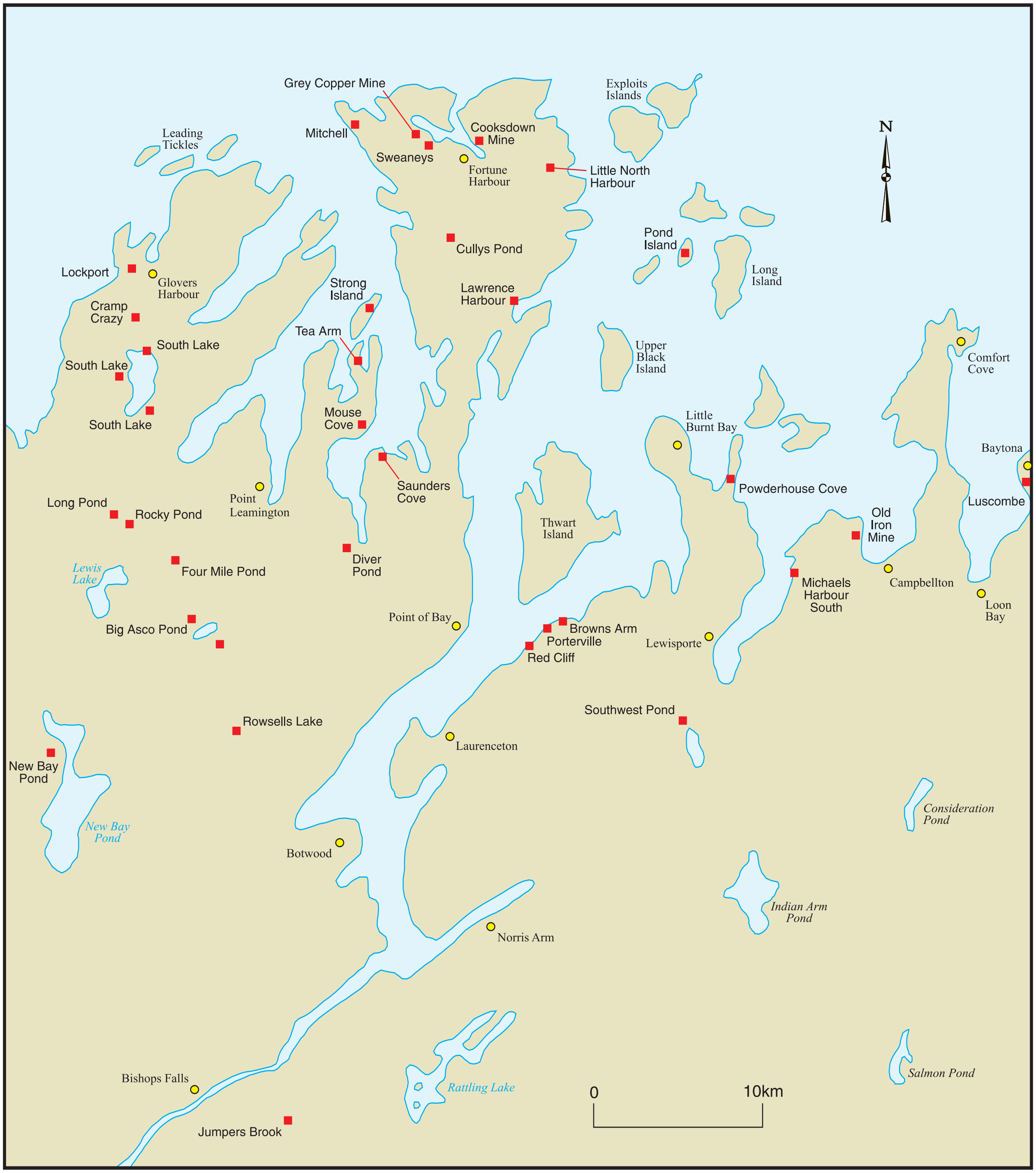


Figure 47. *Geographic location of the known mineral occurrences and prospects in the central Notre Dame Bay region. Modified from MODS-PC (Stapleton and Parsons, 1993).*

Harbour area of New World Island, the cherts and felsic tuffs succeeding the basalts of the Western Head Formation belong to the Haywards Cove Formation of the Moretons Harbour Group. These have been interpreted as the extrusive equivalents of the felsic dykes thought to control the As, Au, Sb, Cu and Zn mineralization at much lower stratigraphic intervals (Kay and Strong, 1983). On the Fortune Harbour Peninsula, there are numerous quartz-feldspar porphyries and composite rhyolite–diabase dykes exposed between Waldron Cove and Little North Harbour.

Western Exploits Group and Eastern Wild Bight Group

In those parts of the Exploits Subzone surveyed for this report, known base-metal occurrences are clustered in several areas and are mostly hosted by certain units of the Exploits and Wild Bight groups. In the western Exploits Group, in the region between Strong Island and Saunders Cove, numerous chalcopyrite–pyrite occurrences occur along strike in chloritic and sericitic alteration zones in several members of the basal Tea Arm Formation. In the eastern Wild Bight Group, reported showings of Cu, Zn, Ni or Ag mineralization are located within more widespread zones of disseminated and stringer pyrite. They occur discontinuously along an arcuate belt of structurally disrupted volcanic, intrusive and sedimentary rocks extending from Glovers Harbour southeastward to Nanny Bag Lake and then southwestward to Rowsells Lake and New Bay Pond.

The most prospective areas for base metals are found in the southwest corner of the study area within a structurally complex tract of variably altered volcanic, volcanosedimentary and intrusive rocks belonging to the Wild Bight Group, the Exploits Group and the Phillips Head igneous complex (Figure 44). Several isolated sulphide showings occur in separate tectonic panels of these Ordovician rocks in the region northwest of the Northern Arm Fault and the Silurian Botwood Group. There, the Ordovician host rocks are located within a large antiformal culmination of a regional thrust stack (Figure 15).

In the Wild Bight Group, zones of disseminated pyrite with traces of copper mineralization have been reported from areas around Big Asco Pond, Nanny Bag Lake, Long Pond and New Bay Pond (Figure 47). These are all hosted by rhyolite flows or occur near rhyolite–basalt contacts. MacLachlan (1998) has postulated that such occurrences reside in fault-bounded tracts of the basal formation of the Wild Bight Group, although previous workers assigned them to a variety of different map units.

The rhyolite and basalt flows near northern New Bay Pond comprise a tectonic fragment of the lower Wild Bight Group (MacLachlan and O'Brien, 1998), are traceable along strike northwestward to the Point Leamington massive sulphide deposit (Swinden and Jenner, 1992) and have also been discovered to extend much farther southeast than previously recognized (Dickson, 1998). The pyritized rock at a showing near Rocky Pond is a gabbro but the country rocks to this mafic intrusion probably also belong to the highly prospective basal unit of the Wild Bight Group. An important exception to these occurrences is the reported chalcopyrite–pyrrhotite mineralization near Rowsells Lake (Figure 47), which is located in a pillow breccia lenticle within the volcaniclastic turbidites of the Pennys Brook Formation of the upper Wild Bight Group.

In the area surveyed, the two most significant areas of known base metal mineralization are the Tea Arm prospect in the Exploits Group near Strong Island Sound and the Lockport prospect in the Wild Bight Group near Glovers Harbour (Figure 47). Both are complexly deformed stockworks containing sulphide pods and stringers but without any significant amount of bedded sulphide exposed at surface. Regionally, the prospects are located in the lowest-exposed parts of the Exploits and Wild Bight groups within comparable intraoceanic suites of early Ordovician arc volcanic and hypabyssal rocks. In particular, the distinctive lithogeochemical fingerprints of refractory arc-tholeiite and high-silica trondhjemitic extrusions adjacent to the mafic dyke-injected alteration zones are very similar (Swinden, 1987; O'Brien *et al*., 1997; MacLachlan and Dunning, 1998a).

Base-metal Prospects in Early Ordovician Rocks

The Tea Arm and Lockport prospects are bounded by major fault zones that, farther southeast along their structural trace, juxtapose early-mid Ordovician arc-derived supracrustal sequences against deep-level and high-level Ordovician plutonic complexes (O'Brien, 1991; O'Brien, 1993; MacLachlan, 1998). The mineralized early Ordovician strata at the Tea Arm and Lockport prospects occur in close spatial proximity to unaltered mid Ordovician volcanic and volcaniclastic rocks which reside in the upper and middle stratigraphic levels of the Exploits and Wild Bight groups. The younger Ordovician rocks are interpreted to have formed in completely different tectonic environments than the older Ordovician rocks (Dec *et al*, 1992; O'Brien *et al*., 1997; MacLachlan and Dunning, 1998b; MacLachlan and O'Brien, 1998).

Tea Arm Prospect

The Tea Arm Cu–Zn prospect is located near the folded and faulted stratigraphic boundary between the arc-tholeiite pillow lavas of the lowest Little Arm East member and the felsic to intermediate pyroclastic rocks of the middle Pushthrough member of the Tea Arm Formation (Figure 13). Smaller alteration zones found farther north on the mainland and Strong Island are locally associated with spectacular epidosites (Plate 48). In the Pushthrough sequence immediately northeast of the prospect, basal agglomerates lying conformably above a scoraceous basalt substrate contain flow-banded and glassy rock fragments but are seen to grade vertically into tuffs rich in embayed crystals. Isotopically dated at 486 ± 3 Ma (Figure 12), the crystal tuffs are lateral-

Plate 48. *Concentric epidote-rich and hematite-rich bands surround fractured relics of vesicular basalt in a stockwork alteration zone within the Tea Arm Formation. Textural modification and mineralogical replacement of the altered basalt preserves the original bulbous shape of the pillowed lava (base outlined by dark lines).*

ly transitional with the dominantly fine-grained pyroclastic rocks (and subordinate felsic breccia) found in the northern part of this member. Dean (1977; 1978) correlated the altered basalt and rhyolite of the Tea Arm Formation with the altered basalt and rhyolite of his Side Harbour Formation of the Wild Bight Group, which he thought hosted the Point Leamington massive sulphide deposit (Walker and Collins, 1988).

The pyroclastic deposits in the Tea Arm area have been interpreted to be vent-proximal and to contain coeval bombs of felsic and mafic magma. Reaction rims around hydrothermally-altered basaltic ejecta remained hot enough to be flattened during compaction and to be distorted during lateral pyroclastic flow. Thinly laminated, jasper-rich chert occurs as a cap rock to the well-developed, marine tuff sequence in Tea Arm bottom; whereas, near the Cu–Zn prospect, it is present as blocks within felsic agglomerate below the top of the Pushthrough section.

In most of the alteration zones between Strong Island and Tea Arm bottom, felsic volcanic strata are observed to be much less altered than the mafic pillow lavas with which they are immediately juxtaposed. The stockwork at the main prospect is apparently hosted by Little Arm East basalts (Plate 49); however, at least in coastal exposures, nearby felsic tuffs of the Pushthrough member are not as strongly affected. According to Swinden (1988) and Swinden *et al*. (1990), base-metal-rich fluids rose along the same conduits and fractures as the magmas which crystallized to form the high-silica trondjemite and the severely LREE-depleted basalt. Moreover, the original sources of these coupled felsic–mafic melts were postulated to be deep in the crust at the mafic base of the island-arc and in hot refractory parts of the underlying mantle wedge, respectively.

Immediately south of the Tea Arm prospect, southeast-dipping, right-way-up felsic tuffs are overlain by unaltered arc tholeiites which are also southeast-dipping and right-way-up. They are geochemically similar to arc tholeiites in the Little Arm East member below the stockwork zone but distinct from the younger refractory tholeiites of the basal Pleasantview member, which occurs farther up the sequence of Tea Arm rocks exposed along the South Arm of New Bay (Figure 13; O'Brien *et al*., 1997). This may indicate interstratification of Little Arm East basalt and Pushthrough rhyolite at a gradational boundary located above the regionally discontinuous felsic tuff horizon. Alternatively, it might imply that a portion of the Little Arm East member, which was originally situated away from the stockwork zone, was structurally repeated within an imbricate fault zone localized near the Pleasantview–Pushthrough contact.

Mafic Intrusions

Ordovician mafic intrusive activity predated, overlapped and postdated the alteration and mineralization seen at the Tea Arm prospect. An extensive regional swarm of diabase and gabbro occurs in the Little Arm East member of the Tea Arm Formation, where the intrusions are observed to be mainly unaltered. However, in a few localities west of Tea Arm (Figure 13), these intrusions and their country rocks are locally pyritized within small alteration zones peripheral to the main stockwork. Some geophysical anomalies in the northern and southern part of the Tea Arm Formation have been interpreted to reflect the presence of mafic to intermediate intrusive bodies at depth (Woolham, 1995). It is possible that the mafic dyke swarm emanated from these inferred subvolcanic magma chambers, as some sheeted amygdaloidal diabases contain coarse-grained plutonic xenoliths.

A younger suite of syn-Pushthrough mafic dykes is localized above the Tea Arm stockwork zone, where the intrusions are seen to be chilled, disrupted and partially fragmented within certain felsic pyroclastic flow sheets. However, on Strong Island, they are observed to pass continuously from the unaltered basalts of the Little Arm East member to the stratigraphically overlying felsic tuff and breccia beds of the Pushthrough member. Minor mafic intrusions emplaced immediately above the Tea Arm stockwork illustrate geochemical evidence of alteration (O'Brien *et al*., 1997); whereas, some of the pyroclastic flows hosting the syn-Pushthrough suite appear to be relatively fresh.

By comparison, the youngest gabbroic intrusions in the Tea Arm area are columnar-jointed sheets which are present in both younger and older parts of the Tea Arm Formation. Banding is developed parallel to the intrusive margin and, in places, this layering is brecciated and cemented by fresh vesicular diabase. This implies rapid cooling and degassing of magma beneath the seafloor at a time when the main pulse of stockwork alteration had presumably ceased. Postdating silicification and pyritization of the Little Arm East pillow lavas, these relatively large unaltered intrusions crosscut the footwall and hangingwall of the prospect and contain xenoliths of altered wall rocks (Plate 50).

Plate 49. *Arc tholeiite lavas in a pyritic and silicic alteration zone near the Tea Arm prospect display relict pillowed shapes. The alteration zone occurs below unaltered felsic pyroclastic rocks of the middle Tea Arm Formation.*

Silicification and Chert Veins

Strongly LREE-depleted, refractory arc tholeiites (TA-3 in O'Brien *et al.*, 1997) occur in the lower Pleasantview member immediately southeast of the Tea Arm prospect (Figure 13). These tholeiites illustrate evidence of pillow collapse, draining and fluid infiltration (Figure 48). To a lesser extent, these features are also observed in the Little Arm East member on islets southeast of Strong Island within a stratigraphical succession of pillowed basalts lying below small unmappable lenses of felsic tuff.

Plate 50. *Unaltered layered tholeiitic intrusion cuts altered pillow breccia and arc-related pillow lava near the Tea Arm prospect.*

Chert veins are observed within the centres of partially collapsed pillows or throughout relics of completely collapsed and more chloritized pillow lavas (Plate 51). They are similar in composition to the variegated interstitial chert seen in the cusp area of pillowed basalts in uncompacted sequences. These chert veins display internal red, green, grey and black laminae formed by repetitive cracking and sealing, and they have smaller pinnate side veins with asymmetrically forked branch terminations. The most intense veining occurs where the basalts are most flattened and where preserved pillow centres contain the most veined and altered metabasite assemblage.

In some flattened pillow lavas, discrete laminae within certain flat-lying chert veins can be observed to extend continuously into adjacent veins that cut steeply across the drained pillows. This indicates that these veins occupied fractures that were open and filled by fluid when the vertical and horizontal fracture surfaces abutted each other. Together, such veins comprise a three-dimensional honeycomb network. It appears that, with increased systematic veining, isolated networks became interconnected and fluid pathways were established between individual sections of collapsed pillow lavas.

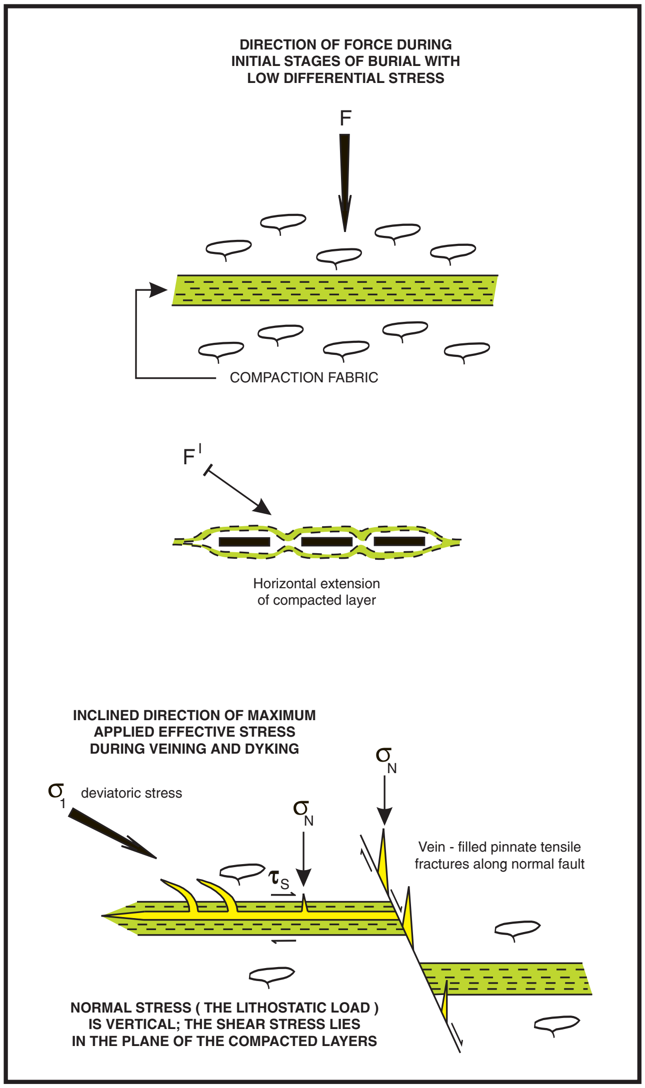


Figure 48. *Model depicting the development of chlorite schist zones with flat, vertical and curved chert veins in flat-lying, differentially compacted sequences of collapsed pillow lava in the Tea Arm Formation.*

Some vertical veins in the chert array were lithified and already solid when they were buckled during loading of the volcanic sequence. Locally, the growth texture of these crack-seal veins was partially destroyed in the recrystallized hinge zones of recumbent microfolds, which are restricted to the collapsed and flattened pillow lava horizons. The formation and distortion of the honeycomb network is indicative of coeval episodes of vertical flattening and subhorizontal shearing, which preferentially affected the most compacted basalt sequences (Figure 48).

Vertical flattening of the basalts produced a flat compaction fabric defined by the preferred optical alignment of neocrystallized penninitic chlorites, presumably in response to the lithostatic stress. In addition to these alteration- related chlorite schist zones (Plate 51), oblate fragments of chloritized pillow basalt, early-formed chert veins and glassy pillow selvages outline a shape fabric parallel to the chlorite schistosity. Subhorizontal shearing along these fabrics is interpreted to have caused the drag folding and lateral offset of the vertical chert veins as well as the boudinage and ductile pinch-and-swell of some horizontal veins (Figure 48). In places, undeformed veinlets crosscut the deformed array of chert veins.

The phenomena described above were contemporaneous with other geological events that affected the Tea Arm lavas during their accumulation. These include quartz veining and bilateral silicification, secondary carbonate precipitation, diabase dyke injection and the increased accumulation of primary interflow limestone. The mafic dykelets, which were emplaced late during the secondary chert-veining process (Plate 52), are presumed to feed lavas at higher stratigraphic levels of the Tea Arm Formation, although geochemical data to support this assertion is lacking.

Siliceous Oxide-facies Iron Formation

The uppermost part of the Pleasantview member of the Tea Arm Formation contains unique horizons of a refractory basaltic andesite, whose LREE patterns resemble those of boninite (O'Brien *et al*., 1997). In the Pleasantview–Saunders Cove transition, the refractory basaltic andesites are interstratified with a red, siliceous, oxide-facies iron formation. In general, these horizons contain a greater abundance of jasper-cemented mafic fragmental rocks in those sections through the transition zone which are located to the west of the Saunders Cove area. Here, and particularly at the Mouse Cove copper showing (Figure 47), the tract of refractory basaltic andesite is reported to host veins and disseminations of pyrite with minor chalcopyrite. Mineralization and alteration have not been observed in the thick Pleasantview succession of low-K tholeiites and calc-alkalic basalts (O'Brien *et al*., 1997) that outcrop between the collapsed refractory basalts near the Tea Arm prospect and the pyritized boninitic lavas associated with the iron formation.

Helwig (1969) considered the iron-oxidized tops of these upper Tea Arm– lower Saunders Cove lavas as indicative of past subaerial conditions and emergence of the early Ordovician arc. However, the writer believes such features are equally likely to be the result of shallow marine (oxic) or sub-seafloor hydrothermal conditions that prevailed during the period of boninitic eruptions. The hydrothermal system affected the red chert and oxide- facies iron formation of the lower Saunders Cove Formation but it probably became dormant during deposition of younger parts of this formation. This is because oxidized boulders of arc basalts similar to those in the Pleasantview–Saunders Cove transition (but different from those encountered in the rest of the Tea Arm Formation) occur as clasts in the succeeding New Bay Formation of the Exploits Group (Hughes and O'Brien, 1994). Although the early-mid Ordovician rocks of the upper New Bay Formation are themselves locally altered at considerable distance from the Tea Arm Formation, they are far too young to be affected by the early Ordovician stockwork mineralization described above.

Lockport Prospect

The Lockport Cu–Zn prospect (Figure 47) occurs southwest of Glovers Harbour in a heavily thrust-faulted anticlinorium which disposes discontinuous portions of the lower, middle and upper parts of the Early-Middle Ordovician Wild Bight Group. To the northeast near Leading Tickles, some of these fault structures dip southeastward and are responsible for placing the Wild Bight Group above Late Ordovician strata assigned to several other rock units (O'Brien, 1991; MacLachlan, 1998). Similar relationships are found southwest of the Lockport prospect near South Lake, except here the thrust faults bounding the Wild Bight Group dip northwestward (Figure 45). Swinden (1987) noted that the setting of this base-metal prospect was more typical of those hosted by felsic and mafic primitive-arc rocks and that elsewhere such lavas lay well below the stratigraphic top of the Wild Bight Group. However, it was not until MacLachlan's (1998) work that the host rocks of the Lockport prospect were correctly assigned to the oldest observable unit in the Wild Bight Group.

In surface exposures of the Lockport prospect, variably silicified pillow lavas occur in a cupriferous stockwork, which has been historically reported to contain minor silver, gold and supergene copper. These LREE-depleted, island-arc tholeiites have the same distinctive lithogeochemistry as the mafic host rocks at other known massive sulphide occurrences in the Wild Bight Group (Swinden, 1987, 1988). Recently, several fault panels of high-silica, low-potassium trondhjemitic rhyolite and quartz–phyric felsic tuff have also been mapped between the Lockport prospect and the South Lake Igneous Complex (MacLachlan, 1998).

Plate 51. *Cross-section of completely collapsed, drained and compacted pillow lava from the Tea Arm Formation with secondary chert veinlets preferentially injected parallel to a horizontal chlorite foliation.*

Plate 52. *Syntaxially-injected diabase dykelets occupy the same conduits as variegated chert veinlets. They document coeval magmatism and alteration, and contemporaneous dilation of horizontal and vertical fractures in the Tea Arm Formation.*

Small alteration zones located northeast and southwest of the Cu–Zn prospect developed in basalt and rhyolite; however, they are disrupted by numerous shear zones and small faults. In the region surrounding the Lockport stockwork, weakly altered lavas occur within lensoid thrust slices. There, fault-bounded volcanic successions alternate from being structurally inverted to being right-way-up and they are observed within a southeast-dipping thrust stack. However, north of the Lockport prospect, the Glovers Harbour volcanic rocks are also seen to comprise right-way-up tectonic panels that dip northwestward.

In the Glovers Harbour–Lockport area, stratiform graded bands in large gabbro bodies yield geopetal information which is in agreement with the stratigraphical facing directions in adjacent pillow lavas. Some gabbro dykes are, however, observed to crosscut flow-banded vesicular diabase sills and concordant screens of pillow lava. Most of these country rocks are unaltered and belong to MacLachlan's (1998) LREE-depleted suite of low-Ti basalts. However, on the coast north of the main Lockport prospect, an unaltered gabbro contains xenoliths of silicified lava and crosscuts completely replaced basalts with relics of pillow structure (Figure 14). According to MacLachlan (1998), this intrusion is coextensive with a gabbro sill having primary alkali basalt chemistry and a U/Pb isotopic age of 486 ± 4 Ma. Thus, the absolute age of arc volcanism and mineralization at Lockport is unknown, although both are demonstrably early Ordovician or older in age.

Relations with Middle Ordovician Rocks

Northwest of the Cu–Zn prospect, in the Locks Harbour area, a northwest-younging sequence of red chert, red debrite, plutonic boulder conglomerate and green turbidite sandstone is interstratified with minor amounts of mafic and felsic tuff. This sequence, which structurally overlies the refractory pillow lavas hosting the Lockport prospect, contains a latest Arenig–earliest Llanvirn felsic tuff dated at 472 ± 3 Ma (Sample 95092 in MacLachlan, 1998). Similar Wild Bight Group strata also crop out within thrust sheets to the immediate southeast of the Lockport prospect. There, presumed latest Arenig and Llanvirn rocks are generally southeast-younging; however, they are disposed in structurally inverted and right-way-up rock sequences adjacent to the Glovers Harbour volcanic rocks. MacLachlan and Dunning (1998b) demonstrated that these unaltered strata comprise part of a regionally developed, early Middle Ordovician succession within the Wild Bight Group and postulated that such strata might have been originally deposited above the older mineralized rocks in the Glovers Harbour tract.

Numerous high-level intrusions of crescumulate gabbro and vesicular diabase are present in dated Middle Ordovician strata near the margins of the younger volcanosedimentary sequence in the Wild Bight Group. In the coastal section east of Goat Island and in the Locks Harbour area, spectacular examples of magma-sediment intermixing are seen along the margins of some of these intrusive bodies. Rare outsized detrital fragments of silicified pillow basalt, similar to those that host the cupriferous stockwork at Lockport, are observed in conglomerate horizons in the same succession. Features that may geologically link the Wild Bight Group's Lockport prospect with the Exploits Group's Tea Arm prospect are the direct juxtaposition of early Ordovician volcanic and mid Ordovician volcano-sedimentary rocks near these Cu–Zn prospects, discrete early and mid Ordovician phases of gabbro and diabase injection near the boundary of the juxtaposed units, and erosion and deposition of variably mineralized early Ordovician magmatic clasts within mid Ordovician and older volcanosedimentary units.

Similar geological relationships are found in less explored regions situated near the northeast and southwest margins of the Wild Bight Group. There, undated Ordovician lavas displaying fundamentally different types of arc-related geochemical signatures crop out in positions that, on regional stratigraphical grounds, would appear to be anomalously high in the Wild Bight stratigraphy (e.g., Williams and O'Brien, 1994). Examples include the northwest-trending structural tract of the Nanny Bag Lake rhyolite, Northern Arm basalt and Phillips Head basalt, and the northwest-trending structural tract of the New Bay Pond rhyolite, Side Harbour basalt and Big Lewis Lake basalt (Swinden and Jenner, 1992; O'Brien, 1993; MacLachlan and O'Brien, 1998; Dickson, 1998).

South Lake and Phillips Head Igneous Complexes

Chalcopyrite mineralization has been reported in sheeted diabase dykes that were intruded into sheared metagabbro (arc ophiolite) in the oldest part of the South Lake Igneous Complex (MacLachlan and Dunning, 1998a). Small copper showings at South Lake and Cramp Crazy Lake (Figure 47) are found in tonalite and quartz diorite, respectively, and they may be unrelated to the showings in the older diabase dyke swarm. Some of the copper mineralization in the South Lake Igneous Complex may be epigenetic and reflect a deeper level of the type of alteration seen in the Moretons Harbour Group on the Fortune Harbour peninsula and the northwestern part of New World Island. Alternatively, the chalcopyrite-bearing sheeted dykes may have been altered in a small subvolcanic stockwork zone and be related to the syngenetic type of copper mineralization (Dean, 1977) observed in the boninite–tholeiite transition zone of the Betts Cove arc ophiolite on the Baie Verte peninsula (Bedard *et al*., 1998).

The mafic plutonic rocks and diabase dykes of the Phillips Head igneous complex are intensely altered and extensively hydrated, although pyrite mineralization appears to be confined to quartz lodes in diorite at the faulted margins of the complex. The mafic volcanic rocks in the Phillips Head igneous complex have arc-related geochemical signals (O'Brien *et al*., 1994); however, they are fresh and not nearly as LREE-depleted as the basalts associated with massive sulphide lenses elsewhere in the Exploits Subzone (Swinden, 1991). The Phillips Head pillow lavas may be prospective in that they possibly correlate with the tectonically adjacent tract of Northern Arm Basalt (Wild Bight Group), which is host to minor Cu–Ag mineralization at the Four Mile Pond alteration zone (Figure 47).

UNITS WITH POTENTIAL FOR SEDIMENT-HOSTED PRECIOUS- AND BASE-METAL DEPOSITS

Base metals, some precious metals and a variety of trace elements are enriched above background sedimentary rock levels in parts of several Middle-Late Ordovician black shale and chert formations in central Notre Dame Bay (Dean and Meyer, 1983). Although some of the underlying Middle Ordovician and older sedimentary units are regionally altered and locally mineralized (e.g., Batterson *et al*., 1998), till geochemistry is lacking where textural and mineralogical evidence of syngenetic alteration is present (e.g., O'Brien and MacDonald, 1997; Rees, 1999). Nevertheless, there is potential for mid Ordovician exhalative-type and replacement-type mineralization in the area surveyed.

Condensed Llandeilo–Caradoc shales were deposited as sulphidic pelagic muds during a regional transgression at the highest Ordovician stand of sea level recorded in the Exploits Subzone. Within the map area, such strata are found in the graptolite-bearing parts of the Shoal Arm Formation (Espenshade, 1937; Dean, 1978), the Lawrence Harbour Formation (Williams, 1995), the Luscombe Formation (Kay, 1975) and/or the Baytona formation (Currie and Williams, 1995). In most localities, the Llandeilo–Caradoc pelagites are separated from latest Arenig–earliest Llanvirn pillow lavas by highly variable thicknesses of ribbon chert and epiclastic wacke, although clastic carbonates or olistostromes locally predominate.

Various colour variations, mottling patterns, nodule types, coticule horizons and secondary cherts are present in the sedimentary successions situated below the main interval of graptolitic mudstone. Such rocks comprise parts of the Exploits Group, the western Dunnage Melange, the eastern Wild Bight Group and the lower Shoal Arm Formation. Northwest of the Dunnage Melange, this late Arenig–early Caradoc sequence is relatively thin, mainly red and green in colour and dominated by chert. Southeast of the Dunnage Melange, it is relatively thick, mostly grey in colour and dominated by siliciclastic turbidites.

Upper Parts of the Exploits Group and Easternmost Wild Bight Group

Red argillite, green chert and grey porcellanite are the dominant rock types in the Strong Island chert of the Exploits Group (O'Brien, 1997) and the uppermost Pennys Brook Formation of the Wild Bight Group (Dean, 1977; Unit 7a of MacLachlan, 1998). Descending these thin-bedded successions of fine-grained to cryptocrystalline rocks, thick pebbly wackes derived from immediately underlying within-plate basalts (Dec *et al*., 1992) become an integral part of the sedimentary sequence. In places, replacement textures associated with the development of secondary prehnite, quartz and hematite (Frank, 1974) affect the argillites, nodular cherts and coarser volcaniclastic turbidites. This type of alteration is observed at even lower stratigraphic levels in the eastern part of the study area, such that strata situated well below the Arenig–Llanvirn pillowed basalts are affected along with the overlying sedimentary sequence (e.g., the altered New Bay Formation east of Burnt Bay).

In the type area of the Lawrence Harbour Formation, Ba-enriched concretions and highly phosphatic fluoroapatite nodules (Dean and Meyer, 1983) have been reported near the grey chert–black shale contact, which is the redefined base of that unit (Williams, 1995). In the red, green and grey chert-dominated divisions of the lower to middle Shoal Arm Formation (Kusky, 1985), the alteration zones are slightly thicker, contain abundant pyrolusite-coated nodular beds and are distributed over a wider stratigraphic interval. Intercalations of grey manganiferous chert and dark grey pyrolusite-rich cherty argillite occur at this level and higher in the Shoal Arm stratigraphic section, where they separate interbanded sequences of turquoise and maroon clastic sedimentary rocks. In Shoal Arm sections near New Bay Pond, Northern Arm and New Bay River, pink-hued podiform coticules are well developed in red laminated siliceous argillites; however, these are interbedded with radiolarian-bearing biogenic cherts (Bruchert *et al*., 1994). Within most mid Ordovician rock units in central Notre Dame Bay, replacement-type cherts are most commonly associated with sedimentary sequences that lie below a grey bioturbated chert with distinctive black shale partings.

In the Osmonton Arm and Leading Tickles areas, some thin-bedded wackes and laminated green argillites are known to belong to the *N. gracilis* biozone (Llandeilo–Caradoc interval) of the uppermost Pennys Brook Formation of the Wild Bight Group. Locally, the entire late Arenig–earliest Caradoc section is less than 500 m thick, is dominated by red chert, and contains very rare alkali basalt tongues (MacLachlan, 1998). A single olistostrome horizon only measures about 2 m in thickness. Pseudoporphyroblasts are locally present where these sedimentary rocks are silicified, carbonatized and pyritized (Plate 53). In contrast, in the Rowsells Pond area west of Northern Arm (Figure 47), several extensive pillow lava lenticles are interstratified with coarse turbidites and olistostromes throughout a 1.5 to 2 km thick section of the Pennys Brook Formation. Here, Pennys Brook turbidites locally display spotted porphyroblasts and hematized and chloritized Pennys Brook basalts contain minor chalcopyrite.

Southeast of the Dunnage Melange, similar tracts of altered turbidites are found within the Strong Island chert in North Campbellton (near the adit location in Figure 16) and also to the south of Campbellton (near the Rod and Gun location in Figure 35). In the vicinity of the Michaels Harbour South copper showing and the Old Iron Mine in Campbellton Harbour (Figure 47), such strata are highly silicified and pyritized, and they contain cherty nodular carbonate lenses. Some of these siliceous greybeds have pink-hued coticules and, farther east in the Luscombe Formation of the Campbellton sequence, similar layers are reported to contain rhodochrosite, rhodonite and other ferromanganese-rich carbonates, silicates and oxides (Kay, 1975). Minor beds of airfall felsic tuff occur low in the altered turbidite section near Indian Arm Brook, Purbecks Cove and Strong Island. Typi-

Plate 53. *Lineated pseudoporphyroblasts of ferroan carbonate and hematized pyrite occur in Osmonton Arm in selected horizons of thin-bedded siliceous turbidites belonging to the Pennys Brook Formation of the Wild Bight Group.*

cally, the felsic tuffs are a few centimetres thick and they lie close to the underlying late Arenig–early Llanvirn pillow lavas of the Lawrence Head Formation (O'Brien *et al*., 1997).

The most extensive sub-Lawrence Head tract of stratabound alteration in the Exploits Group widens eastward from Sand Cove in New Bay through the Winter Tickle–High Grego Island area and onto the northwest coast of Upper Black Island. Here, numerous differentially compacted intervals of sericitized and albitized wacke in the Saltwater Pond member illustrate honeycombed arrays of quartz-carbonate veins. This same stratigraphy is purported to have been reworked in the Dunnage Melange (Hibbard and Williams, 1979). The diagnostic coticule horizons of the unbroken formations of this melange have been interpreted to have formed in manganese-rich sediment deposited in the deep-sea environment within restricted sedimentary basins (Williams, 1992).

Biozones, Lithofacies and Sulphide Mineralization

Throughout most of the central Notre Dame Bay region, the Caradocian time interval–the *N. gracilis, C. bicornis* and *D. clingani* graptolite biozones–is mainly represented by rocks of the black shale lithofacies. By way of contrast, in Llandeilo and Ashgill times, the chert and limestone lithofacies were widely developed in the siliciclastic and calcareous turbidite basins, and the pyritic black shale facies is very rare.

The stratigraphical boundary of Caradocian black shale with older sedimentary rocks is locally preserved near the top of the Wild Bight and Exploits groups, and coeval black shales are purported to have originally covered the top of the Dunnage Melange and the Summerford Formation (Hibbard and Williams, 1979). The stratigraphic base of the black graptolitic shales can be observed within the Luscombe Formation near Campbellton, the Shoal Arm Formation near New Bay Pond and the Lawrence Harbour Formation near Osmonton Arm bottom. However, in at least parts of these formations, development of the black shale lithofacies was apparently restricted to mid-to-late Caradoc time, as most shales in question lie in the *D. clingani* Zone with only small tracts of the *C. bicornis* Zone having been identified (S.H. Williams, 1991, unpublished data; Williams, 1993). In places, where the uppermost bioturbated mottled cherts reside in the *C. bicornis Zone* (Williams and O'Brien, 1994), the black shale lithofacies did not accumulate in the early to mid Caradoc interval, and the underlying silicified turbidites are known or presumed to lie in the *N. gracilis* Zone and older graptolite zones of the Middle Ordovician (Table 1).

Northwest of the town of Point Leamington, on Western Arm Brook and Mill Pond (O'Brien, 1991), rhythmically interbedded light grey mudstone and dark grey nodular argillite occur directly below *C. bicornis* Zone and *D. clingani* Zone shales. At Campbellton's Old Iron Mine (Figure 18), similar strata underlie *D. clingani* Zone black shales which are marked by conspicuous pyrite-rich and pyrrhotite-rich layers (A. Sangster, personal communication, 1992; Williams, 1993). In several of these localities, debris flow deposits interbedded with the silicified turbidites contain relatively unaltered blocks of fossiliferous limestone (O'Brien, 1992a). The lack of limestone dissolution may indicate that silicification of these 'stockwork' turbidites occurred at shallow levels beneath the sea floor and that silica precipitation was probably controlled by through-going fluids pressurized by a carbon dioxide gas (Hesse, 1990).

Sangster (1993) demonstrated that, within the Luscombe succession (Table 1; Figure 47), gold concentrations attained maximum values in the most reduced part of the sulphide-bearing carbonaceous shale sequence. The geochemical paleoenvironment of the distinctive manganese-rich grey cherts and silicic wackes was interpreted to be mildly reducing; however, more reduced black shales and pyritiferous siliceous argillites occur stratigraphically above this interval. Sangster (1993) concluded that the most auriferous strata are organic carbon-rich, strongly reduced black shales lying closer to the top of this Caradoc succession and that the mineralizing fluids carrying secondary silica and sulphur were partly marine and partly hydrothermal in origin.

Mineralization in the Regional Geological Framework

The sequence of grey biogenic chert, barium-rich nodular argillite and black pyritiferous shale may have formed an impermeable cap rock above the silifiicified and replaced 'stockwork' turbidites at a time when the circulating fluids changed from being oxidized to being reduced. However, if once present, this seal must have been subsequently breached to allow fluid discharge related to the mineralization of the *D. clingani* pyrrhotite beds and possibly, in certain regions, the stratabound silicification, pyritization and carbonatization of early Ashgill turbidites (e.g., Unit 11 of O'Brien, 1992). The phosphatic concretions at Lawrence Harbour could have formed by direct precipitation from seawater, but Llandeilo–Caradoc replacement of nodular limestone is also possible.

In places, there was probably enough Ordovician relief on the sea floor to enable the Llanvirn–Llandeilo Hummock Island limestone to build-up adjacent to age-equivalent chert sequences in the Exploits Group (O'Brien *et al*., 1997). The late Middle Ordovician limestone, grey chert and black shale lithofacies of central Notre Dame Bay were possibly originally disposed across the depositional strike from one another from the margin to the depocentre of a regional black shale basin (e.g., Wignall, 1991). The vertical stacking of such lithofacies in the stratigraphic succession of central Notre Dame Bay presumably reflects the rise of sea level during a late Llandeilo–Caradoc regional transgression. Grey chert–black shale depocentres could be second-order features of the wider basin covering the Exploits Subzone or, perhaps, the relics of older back-arc or arc rift basins (e.g., Bruchert *et al*., 1994). In this regard, the tectonic hinge line situated near the Dunnage Melange may not only have controlled mid Ordovician and subsequent sedimentation (Oversby, 1967) but it may have also been a site on the ocean floor where fluids discharged over the same time interval.

UNITS WITH POTENTIAL FOR EPIGENETIC PRECIOUS-METAL DEPOSITS

Most reported gold occurrences in central Notre Dame Bay are associated with altered rock units of Ordovician and Silurian age. In numerous locations, sheared mafic intrusive rocks are carbonatized and pyritized within certain parts of the intrusive bodies. Commonly, sedimentary rocks are sericitized and silicified along certain segments of regional vein systems. These types of alteration zones are widely distributed throughout the region, although only some of them are known to contain gold mineralization.

Evans (1993) interpreted most of the auriferous quartz vein occurrences in the eastern Dunnage Zone as epigenetic mesothermal deposits. These are hosted by regionally deformed sedimentary and hypabyssal rocks in the suprastructure of the foldbelt; however, the carbon dioxide-rich mineralizing fluids are postulated to have originated in devolatized parts of the metamorphic infrastructure (e.g., Churchill *et al*., 1993). An epigenetic group of epithermal gold deposits (Evans, 1996) were considered to have formed later in the geological history of central Notre Dame Bay. These were noted to occur along lineaments near the margins of Siluro-Devonian plutons and their fractured country rocks (e.g., Tallman and Evans, 1994). The regional boundary faults separating the Botwood, Indian Islands and Davidsville groups have proven to be the most economically significant structures in the eastern Dunnage Zone (Churchill, 1994).

Gold-Related Alteration in the Exploits Subzone of Central Notre Dame Bay

In the area mapped for this report, the mineralized showings generally fall into two types. In the first type, the mineralization appears to be structurally-controlled by ductile shears as the associated alteration zones are commonly located in fault-imbricated units of Ordovician and Silurian stratified rocks (Figures 18 and 47). Typically, pyrite and arsenopyrite with or without chalcopyrite are found in wall-rock alteration zones and in minor veins (e.g., Cullys Pond, Jumpers Brook and Michaels Harbour South). This type of mineralization is also developed in sheared gabbro sills (Plates 54 and 55). Such gabbros are intruded into sedimentary strata which may also be locally altered (e.g., Porterville).

The second type of showing is hosted by late- to post-tectonic intrusive rocks or their adjacent country rocks (Figure 18). Both are heavily jointed and locally offset by brittle fault structures. The gold mineralization is reported to occur in quartz–carbonate–sulphide–oxide veins commonly in association with stibnite (e.g., Pond Island and Powderhouse Cove).

Sulphide Mineralization and Gold Showings near Faults

Southeast of Cottrells Cove at Cuddys Pond, copper mineralization has been reported from highly-sheared basalt olistoliths in the Boones Point Complex adjacent to the fault contact with the Badger Group. This showing occurs in the regional Lukes Arm–Sops Head fault zone. Chalcopyrite is also reported from highly-sheared terrestrial basalt of the Botwood Group south of Michaels Harbour at its thrust-faulted boundary with the New Bay Formation of the middle Exploits Group.

Plate 54. *A variably sheared gabbro sill (intruding hornfelsed rocks of the Badger and Exploits groups) displays a southeast-dipping systematic vein array of bifurcating quartz–carbonate lodes on eastern Upper Black Island.*

Plate 55. *Bilateral alteration zones (marked by a relative abundance of pyrite and ferroan carbonate) occur along the margins of linked quartz–carbonate veins in host gabbro.*

Precious metal mineralization has been reported from the Ordovician–Silurian turbidites of the Badger Group on Jumpers Brook, where the host rocks have been intruded by numerous mafic dykes. The gold zone occurs in an imbricate fault zone that also involves terrestrial basalts and is regionally located at the margin of the Silurian Botwood Group. Southwest of Campbellton, abundant arsenopyrite is seen within pyrolusite-stained cherts of the upper Exploits Group, where these Middle Ordovician strata are thrust faulted above the Botwood and Badger groups.

All of the above units are locally crosscut by felsic porphyries of the Loon Bay suite and, in many of these locations, the porphyries are netveined and pyritized (Figure 18). Loon Bay felsic porphyries are reported to be auriferous in Powderhouse Cove, where they are observed to be drag folded and offset by brittle fractures. Similar quartz feldspar porphyries are also highly altered in several other locations along the faulted margin of the Dunnage Melange.

Extensive pyrite and arsenopyrite mineralization was observed in Ordovician gabbro sills and adjacent turbidites of the Exploits Group near Diver Pond in the bottom of the South Arm of New Bay (Figure 47). This occurrence lies close to a folded fault zone which was injected by sheets of Loon Bay-type granite porphyry.

One of the largest known pyritic and sericitic alteration zones in the map area is located south of Lewisporte within highly-sheared mafic and felsic volcanic rocks of the Botwood Group. The Southwest Pond alteration zone (Dickson and Colman-Sadd, 1993) occurs near the thrust-faulted contact of the Botwood Group with the turbidites of the Badger Group (Figures 18 and 47). Along strike, at Browns Arm, Porterville and Red Cliff, gold mineralization is present within shear zones in sheeted gabbros, although the age of the host intrusions is locally unknown. These localities are situated near a folded fault zone which separates Ordovician chert and basalt (of the upper Exploits Group) from Silurian sandstone and basalt (of the upper Botwood Group).

On Long, Little Berry, Pond, Tinker, Swan and Hornet islands, northeast-trending fractures, closely spaced joints and small faults extend from the northern margin of the Loon Bay batholith, traversing both the plutonic rocks and the melange tracts of the Red Indian Line imbricate fault zone (*see* Posttectonic Major Intrusions, page 112). Felsite dykes and microporphyries occupy some of these fractures. On Pond Island, these satellite intrusions cut a coarse-grained granodiorite which is strongly altered and heavily jointed. Locally, the porphyries have been reported to contain tetrahedrite and bismuthinite with minor concentrations of Sb and Ag and trace amounts of Cu, Fe, Ni and Zn.

Mineralization in the Regional Geological Framework

In Strong Island Sound and the South Arm of New Bay, some diorite bodies hosted by the Exploits Group were steeply dipping, sheared and selectively altered by ferroan carbonate before other unaltered diabases were intruded along the same conduit. A similar situation exists at Purbeck Cove near Porterville, where gabbros are deformed and altered in narrow shear zones and then intruded by fresh porphyritic diabase. In Norris Arm bottom, New Bay Pond, Campbellton and Upper Black Island, deformed gabbro sheets that intrude fossil-bearing Ashgill and Llandovery turbidites are also locally observed to be silicified, ankeritized and pyritized. However, in places, these Silurian intrusions appear to be less altered than their quartz-veined, pyritized and strongly hematized country rocks.

In the area surveyed, the most favourable regional setting for epigenetic deposits is potentially found where the Botwood and Badger groups are regionally thrust-imbricated near the Dunnage Melange or its equivalents. Analogous structures may be present at the gold prospects along the Dog Bay Line (Williams, 1993; Currie, 1993). Syndepositional faults along the margins of the Botwood and Badger basins may have been reactivated as ductile thrust faults in the Late Silurian. At this time, Ordovician and Silurian gabbros behaved relatively stiffly in comparison to the adjacent strata, and the resultant fracture dilatancy focused fluid flow in the ductile shear zones (Churchill, 1994). Rapidly uplifted into near-surface levels of the crust, these gold-bearing rocks had the potential to be re-mineralized during the epizonal intrusion of the Devonian granite porphyries and the fracture-related swarms of plagioclase porphyritic diabase dykes (Figure 17).

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