



TECTONIC IMPLICATIONS OF EARLY SILURIAN THRUST IMBRICATION OF THE NORTHERN EXPLOITS SUBZONE, CENTRAL NEWFOUNDLAND

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Abstract—Central Newfoundland's Dunnage Zone contains a composite assemblage of island arc and oceanic rocks formed in the Paleozoic Iapetus Ocean. Two different and perhaps unrelated Late Cambrian–Middle Ordovician volcanic belts (Lush's Bight and Robert's Arm belts, and their correlatives) are preserved in the Notre Dame Subzone, and are interpreted to represent arc and back-arc basin sequences developed; (1) adjacent to the Laurentian margin of Iapetus and (2) in the intraoceanic realm of the 4000 km wide Paleozoic ocean. Paleozoic paleogeographic reconstructions suggest that the Exploits Subzone contains volcanic and sedimentary rocks originally deposited in a third arc located between the intraoceanic arc (Robert's Arm Belt) and the Peruvian promontory of Gondwana. The New Bay Pond area is located within the northern Exploits Subzone, and preserves an imbricate thrust stack formed during Late Ordovician/Early Silurian southeastward-directed thrusting. Thrust-load related subsidence led to deposition of cherts and argillites overlain by a southeastward-prograding wedge of orogenic flysch and molasse in a foreland trough. Late Ordovician–Early Silurian structural disruption of the flysch/molasse sequence and older strata of the northern Exploits Subzone, by a series of thrust faults, place Middle Ordovician rocks over the younger flysch, and formed a new, southeast-vergent accretionary wedge. This event records the emplacement of the Exploits Subzone over Gondwanan continental margin rocks of the Gander Zone, soon after the Notre Dame Subzone collided with the Appalachian margin of Laurentia. Continued closure of Iapetus brought together the collisionally modified margins of Gondwana and Laurentia. The Exploits Subzone is cut by numerous Silurian–mid Devonian predominantly dextral strike-slip faults, suggesting that this convergence involved a significant component of dextral transpression. Several pull-apart basins are located along these strike-slip faults, and are filled by Silurian volcano-sedimentary sequences. These faults formed in response to the counter-clockwise rotation of Gondwana, juxtaposing the Avalonian margin of Gondwana with the Appalachian margin of Laurentia by the Carboniferous. Copyright © 1996 Elsevier Science Ltd

INTRODUCTION

Northern Newfoundland offers one of the most complete sections across the Appalachian orogen. The complex structural/lithological assemblage preserved in the Newfoundland Appalachians is widely interpreted to result from closure of a complex oceanic basin (Iapetus) similar to the present-day south west Pacific, ultimately bringing together the continental

margins of Laurentia and Avalonia/Gondwana (Bird and Dewey, 1970; Harland and Gayer, 1972; McKerrow and Ziegler, 1972; Dewey *et al.*, 1983; Harris and Fettes, 1988; McKerrow, 1988; van der Voo, 1988; van der Voo *et al.*, 1991; Dunning *et al.*, 1990, 1991; van der Pluijm *et al.*, 1993). In traditional models for the Appalachian orogen, three major orogenic pulses (Taconic, Acadian, and Alleghanian) and two localized events (Penobscottian and Salinic) are interpreted to be associated with accretion of different arc and continental fragments. The Penobscottian event (Neuman, 1967; Boone and Boudette, 1989; Neuman and Max, 1989; Boone *et al.*, 1989; Coleman-Sadd *et al.*, 1992; Chow, 1994) is recognized as distinct deformation locally, from Maine to Newfoundland, associated with ophiolite obduction onto rocks deposited on the Gondwanan continental margin. The Middle Ordovician Taconic orogeny is indicated by ophiolite obduction in western Newfoundland, and deformation of the Cambrian/Ordovician arc rocks in central Newfoundland (Williams, 1979; Williams and Hatcher, 1983; Dewey *et al.*, 1983). The Taconic orogeny is classically interpreted as a collision between the Appalachian margin of Laurentia and a composite island arc terrane (Rodgers, 1968; Rowley and Kidd, 1981; Zen, 1983; Stanley and Ratcliffe, 1985; Bradley and Kusky, 1986). Silurian orogenic activity in the Newfoundland Appalachians includes abundant strike-slip faulting (with pull-apart basin formation), penetrative deformation, metamorphism, and plutonic and subaerial volcanic activity (Karlstrom *et al.*, 1982; van der Pluijm, 1986; van der Pluijm *et al.*, 1993; Kusky *et al.*, 1987; Dunning *et al.*, 1990; Elliott *et al.*, 1991). Late Silurian events have been referred to by some as the Salinic disturbance (Boucot, 1962; Dunning *et al.*, 1990). Devonian plutonic, metamorphic and structural events in the Newfoundland Appalachians are traditionally assigned to the Acadian orogeny, interpreted by most workers to represent continent–continent collision between Laurentia and the Avalonian segment of Gondwana (Rodgers, 1968; Zen, 1983; Bradley, 1983; Dewey *et al.*, 1983; Hatcher, 1988; Roy and Skehan, 1993). Subsequent Permian–Carboniferous events are assigned to the Alleghanian orogeny (Williams and Hatcher, 1983; Rast and Skehan, 1993), largely representing continued convergence including rotation between Gondwana and Laurentia, isolating different Gondwanan continental fragments along the orogen (C. van Staal, pers. comm., 1995).

Considerable controversy has arisen about the relative position in Iapetus of different volcanic belts of central Newfoundland's Dunnage Zone, relative to different margins of the Laurentian and Gondwanan continents during their development, and also about the timing and duration of magmatism in each of these volcanic belts. Traditional models of Appalachian orogenesis have recently been challenged by paleomagnetic, paleontologic, geochronologic and structural data indicating a more complex sequence of events, in which several different arc sequences developed at the same time but in different places, and were later juxtaposed during closure of the Iapetus ocean (van der Pluijm, 1986; Dunning *et al.*, 1991; van der Pluijm *et al.*, 1993; de Roo and van Staal, 1994; van Staal and de Roo, 1996; Dalziel *et al.*, 1994). In this study, we examine the structural setting of a group of volcanic rocks in the Exploits Subzone of central Newfoundland's Dunnage Zone that have proven critical in determining the ages of magmatism and the relative positions of the different volcanic sequences of central Newfoundland. We make regional correlations between this group and others of central Newfoundland, and discuss the implications for models of Appalachian orogenesis.

REGIONAL GEOLOGY

The Newfoundland Appalachians are divided into several tectonostratigraphic zones (Fig. 1; (Williams, 1979; Williams *et al.*, 1988; Dewey *et al.*, 1983)). In the west, the Humber Zone

consists of Grenvillian gneiss representing basement to the Appalachian margin of Laurentia, overlain by a Cambrian–Ordovician platform sequence, and ophiolitic nappes. The Dunnage Zone contains vestiges of the formerly 4000 km wide (van der Pluijm *et al.*, 1990, 1993) Paleozoic Iapetus Ocean, and is subdivided into the Notre Dame and Exploits Subzones (Dewey *et al.*, 1983; van der Pluijm and van Staal, 1988; Williams, 1979; Williams *et al.*, 1988). In eastern Newfoundland, Gander Zone rocks (Gander Lake Subzone) include a continental margin sedimentary sequence now resting adjacent to Precambrian gneiss of the Avalon Zone. Windows of the Gander Zone are exposed through the Exploits Subzone as the Meelpaeg and Mount Cormack Subzones (Fig. 1). Paleomagnetic and paleogeographic data indicate that in the Early–Middle Ordovician, the Appalachian margin of Laurentia was oriented roughly E–W near 11°S, and that the Avalonian continental margin of Gondwana was located near 45–55°S for much of

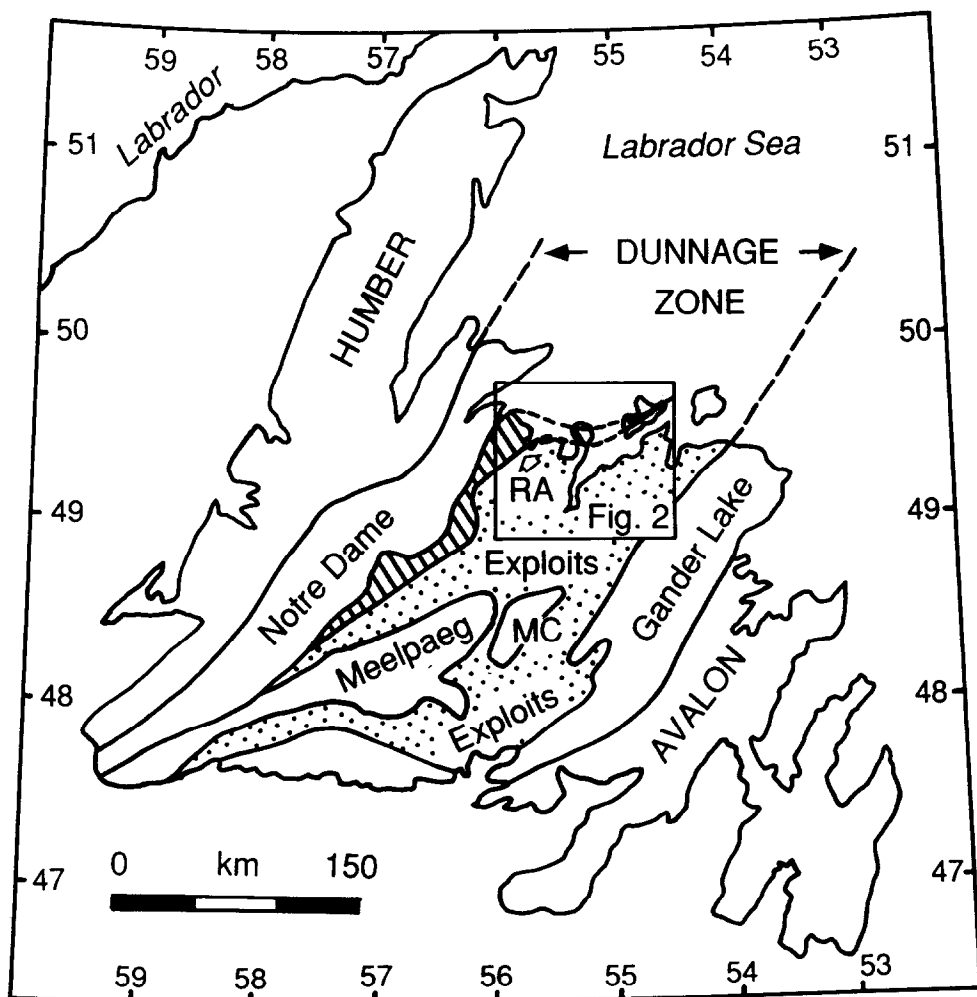


Fig. 1. Map of the tectonostratigraphic zones of Newfoundland (modified after Williams *et al.*, 1988). The Dunnage Zone is divided into the Notre Dame Subzone (including the Roberts Arm Belt (RA)), and the Exploits Subzone, whereas the Gander Zone is subdivided into the Gander Lake, Meelpaeg, and Mount Cormack (MC) Subzones.

the Ordovician (van der Voo, 1988; van der Voo *et al.*, 1991; van der Pluijm *et al.*, 1993).

The Notre Dame Subzone (Figs 1 and 2) contains volcanic belts with Cambrian through Middle Ordovician ages. The Lush's Bight Group and its correlatives (e.g. Lake Ambrose, Moreton's Harbour, Sleepy Cove—Twillingate, and the Baie Verte Peninsula ophiolites), are thought to represent an island arc(s) and/or back-arc basins (Williams *et al.*, 1976; Dewey *et al.*, 1983; Bostock, 1988; Swinden, 1990; Dunning *et al.*, 1991) developed near the Appalachian margin of Laurentia, and colliding with this margin during the Middle Ordovician Taconic orogeny. Paleomagnetic data from these rocks (van der Voo, 1988; van der Voo *et al.*, 1991; van der Pluijm *et al.*, 1993) indicate a paleolatitude of circa 11° S, similar to the E–W trending Appalachian margin of Laurentia. The Robert's Arm Belt (Fig. 2) and its correlatives (including the Buchans, Cottrell's Cove and Chanceport Groups) are part of the Notre Dame Subzone, but

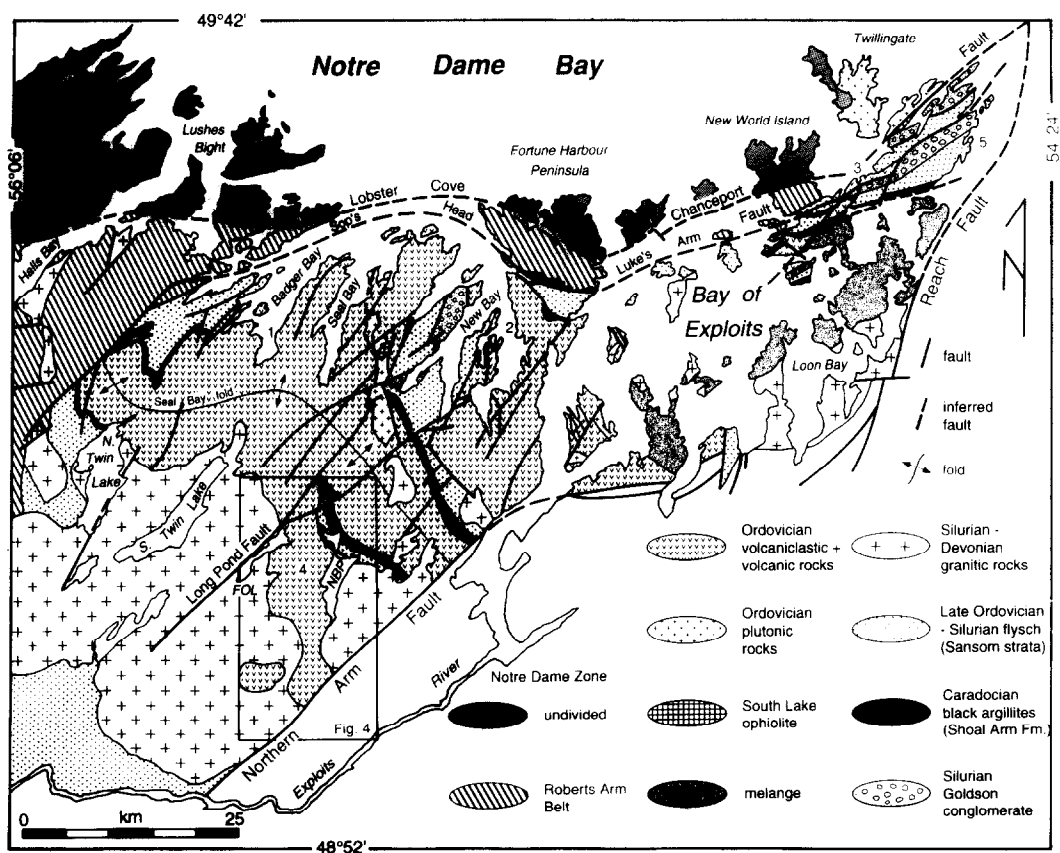


Fig. 2. Generalized geologic map of the northern Dunnage Zone of central Newfoundland, showing the location of the New Bay Pond map area (compiled and modified after Noranda, 1972; Dean, 1978; Kean and Strong, 1980; Kusky, 1985). The Exploits Subzone is bounded on the north and west by a thrust fault (locally known as the Sop's Head, Luke's Arm Fault), and by a dextral strike-slip fault system on the southeast (Reach - Northern Arm - Noels Pauls Faults). Numbers show locations of stratigraphic columns in Fig. 3.

yield significantly different paleolatitudes—30–33° S (van der Voo, 1988; van der Voo *et al.*, 1991; van der Pluijm *et al.*, 1993) and locally preserve Celtic fauna indicative of an intraoceanic island or seamount environment (Neuman, 1984; Jacobi and Wasowski, 1985). The Exploits Subzone (including the Wild Bight, New Bay, Exploits and Summerford Groups) contains an assemblage of volcanic and clastic rocks formed somewhere between these arc-related and intraoceanic island sequences of the Robert's Arm Belt and Gander Zone rocks deposited on the Gondwanan continental margin. The former position of Exploits Subzone strata relative to the Robert's Arm Belt, Gander Zone continental margin sediments, and various continental margins of Gondwana is not well known. Here, we re-examine the stratigraphic and structural development of a critical part of the northern Exploits Subzone, and discuss the implications of these data for the position and tectonic development of the Exploits Subzone within the Iapetus Ocean and the Appalachian orogen.

Figure 2 is a geologic map of the north-central Dunnage Zone, showing relationships between major geologic units and structures. The Notre Dame/Exploits Subzone boundary is marked by the Sop's Head–Luke's Arm Fault (correlative with the Red Indian Line of Williams *et al.*, 1988), whereas the Lobster Cove–Chanceport Fault separates the Robert's Arm volcanic belt (including the Buchans, Robert's Arm, Cottrell's Cove and Chanceport groups) from other parts of the Notre Dame Subzone (Nelson, 1981; La France, 1989; Dewey *et al.*, 1983). Johnson *et al.* (1991) and van der Voo *et al.* (1991) have determined significantly different paleomagnetic poles from rocks on either side of the Lobster Cove–Chanceport Fault, suggesting that the Notre Dame Subzone consists of two or more separate island arc systems. Early–Middle Ordovician mafic and felsic volcanic rocks are widespread throughout the Exploits Subzone, and these are almost invariably capped by Caradocian black argillites and cherts of the Shoal Arm Formation (e.g. Sections 1–5, Figs 2 and 3), indicating cessation of volcanism in Middle Ordovician time. Late Ordovician through Early Silurian, uplift and erosion of a terrane to the northwest is recorded by a younger, upward-coarsening flysch-type sedimentary sequence (Fig. 3) including rocks of the Goldson conglomerate and Sansom equivalent strata (Point Leamington Formation of Williams, 1991). Volcano-sedimentary rocks in the New Bay Pond area (Fig. 4) are enigmatic in that they structurally overlie Caradocian and younger sedimentary rocks (Dean, 1977, 1978; Dean and Kean, 1980; Kusky, 1985; Swinden, 1988), thus preserving clues to post–Middle Ordovician events in the Newfoundland Appalachians.

The Twin Lakes Diorite complex of the prominent Hodges Hill/Twin Lakes pluton in the northern Exploits Subzone has yielded an Early Silurian U–Pb age of 435 ± 2 Ma (G. R. Dunning, pers. comm., 1993), whereas similar plutons of the Mount Peyton pluton have yielded Late Silurian–Early Devonian ages including 420 ± 8 Ma for the Mt. Peyton gabbro (Ar/Ar; Reynolds *et al.*, 1981), 390 ± 15 Ma for the Mt. Peyton granite (Rb/Sr, Bell *et al.*, 1977), 380 ± 16 Ma for the Fogo Island pluton (K/Ar, Wanless *et al.*, 1964), and 408 ± 2 Ma for the Loon Bay batholith (U–Pb, Elliott *et al.*, 1991). The Northern Arm–Reach Fault system cuts through the Exploits Subzone and controls the distribution of several Silurian pull-apart basins (Kusky *et al.*, 1987). Northeast striking, generally dextral, strike-slip faults are common in the area, and Kusky *et al.*, 1987) suggested that these belong to two generations: one preceding intrusion of the Hodges Hill batholith, and one post-dating intrusion. Alternatively, the Hodges Hill batholith may have intruded during a continuous interval of dextral strike-slip faulting. Elliot *et al.* (1991) suggested that the Reach Fault experienced sinistral slip in Ludlow (mid-Silurian times). The sense of movement on orogen-parallel strike-slip faults like the Northern Arm Fault has been controversial, with some workers favoring sinistral slip (Hanmer, 1981; Currie and Piasecki, 1989), others favoring dextral movement (Bradley, 1983; Kusky *et al.*, 1987; van Staal and

Williams, 1988; Lafrance and Williams, 1992), reverse (Cousineau and Tremblay, 1993), or more complex scenarios (Hibbard, 1994; de Roo and van Staal, 1994).

STRATIGRAPHY OF THE FROZEN OCEAN LAKE/NEW BAY POND AREA

Figure 4 is a geologic map of the Frozen Ocean Lake–New Bay Pond area (after Kusky, 1985). Early (?) through Middle Ordovician rocks of the Wild Bight Group, Shoal Arm Formation, and Point Leamington Formation crop out in the northern part of the area, whereas a second outcrop area of volcanic and volcanoclastic rocks of the Wild Bight Group outcrops structurally above the Point Leamington Formation, in the southern part of the area. We informally refer to this southern outcrop belt as the New Bay Pond section of the Wild Bight Group. Early workers (Dean, 1978) described these rocks as a separate and distinct Middle Ordovician–Early Silurian group of rocks (Frozen Ocean Group) in the Exploits Subzone, and correlated them with the Robert's Arm Group of the Notre Dame Subzone (Fig. 2). This correlation however, was discounted by Kusky (1985), Kusky and Kidd (1985), and Swinden (1988). Figure 5 compares the stratigraphic section of the Wild Bight Group on New Bay Pond with a section of the Wild Bight Group in the Badger Bay–Seal Bay area (Fig. 2) mapped by

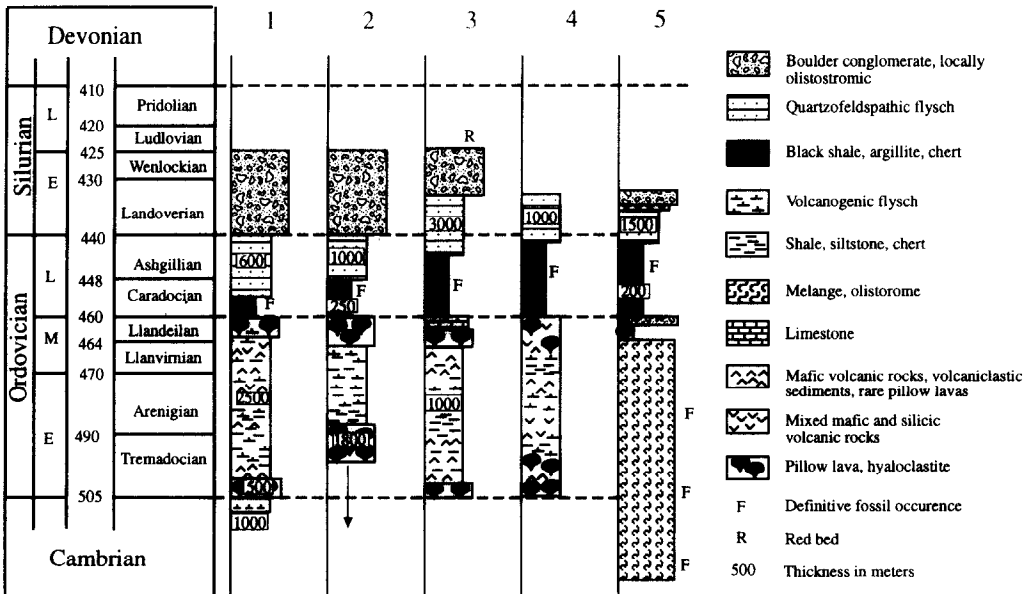


Fig. 3. Stratigraphic correlation chart for the northern Exploits Subzone, after revising the correlation of the New Bay Pond section of the Wild Bight Group with other parts of the Wild Bight Group. After Dewey *et al.* (1983) and Kusky (1985). See also Williams, 1991 and Colman-Sadd *et al.*, 1992.

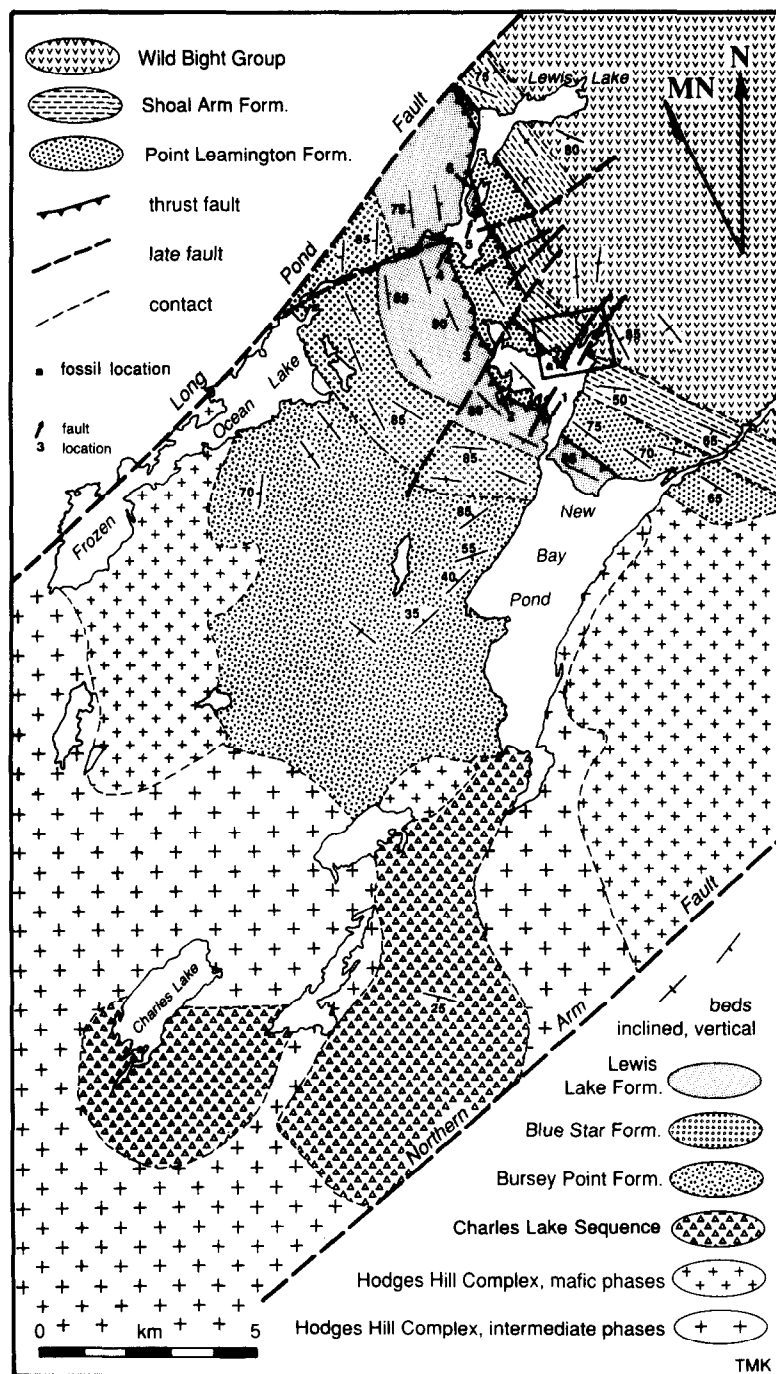


Fig. 4. Generalized geologic map of the Frozen Ocean Lake — New Bay Pond area (after Kusky, 1985). Inset shows location of Fig. 7.

Nelson (1979), showing the remarkable similarity between the two different sequences, and supporting their correlation.

The Early Silurian Botwood Group (Williams, 1962, 1967; Kusky *et al.*, 1987) outcrops to the southeast of the Northern Arm Fault, which cuts across the southeastern end of the map area. Plutonic rocks of the Twin Lakes/Hodges Hill complex intrude all rocks but the Botwood Group,

**Frozen Ocean Group
New Bay Pond**

<u>Burse Point Fm.</u> mafic pillow lava volcanic breccia volcaniclastics (>3.0 km)
<u>Blue Star Fm.</u> volc. sandstone. conglomerate. (1.2 km)
green volc. sst. and argillite
red chert (0.6 km)
<u>Lewis Lake Fm.</u> mixed mafic and silicic flows. pillow lavas. volcaniclastic sandstones (1.7 km)

**Wild Bight Group
Badger Bay - Seal Bay**

<u>Seal Bay Head Fm.</u> mafic pillow lava volcanic breccia volcaniclastics (2.2 km)
<u>Little Harbour Fm.</u> volc. conglomerate sandstone (1.2 km)
red and green volc. sst. and argillite
red chert (0.6 km)
<u>Corner Point Fm.</u> mixed silicic volcanics and mafic pillows. volcaniclastic sandstones (1.9 km)
<u>Omega Point Fm.</u> red and green volc. sandstone and argillite (1.0 km)

Fog Fault

Fig. 5. Comparison of stratigraphy of the New Bay Pond area and the Badger Bay/Seal Bay areas. New Bay Pond section after Kusky (1985), and Badger Bay-Seal Bay section after Nelson (1979).

which is intruded by the similar Mount Peyton Pluton and 422 ± 2 Ma dikes (U-Pb, Elliott *et al.*, 1991).

Wild Bight Group (New Bay Pond section)

Rocks outcropping between New Bay Pond and Frozen Ocean Lake (Fig. 4) include volcanoclastic sedimentary rocks, mafic, intermediate and silicic volcanic rocks, and less-common cherts. Sedimentary and volcanoclastic rocks are more abundant than volcanic rocks, a feature typical of most pre-Caradocian sequences in the Exploits Subzone. These rocks were divided into several northwest trending belts by Noranda Exploration Company (unpublished maps, 1972), and subsequently published in maps compiled by Dean (1977, 1978), who named them the Frozen Ocean Group. The area was remapped by Kusky (1985), and the rocks were correlated with the Wild Bight Group to the north. Swinden (1988) geochemically corroborated this correlation, and suggested that the name Frozen Ocean Group be dropped.

We here informally divide the New Bay Pond section of the Wild Bight Group into four formations. The Lewis Lake Formation forms the structural base of the New Bay Pond section of the Wild Bight Group and is characterized by abundant pillowed mafic lavas. The entirely sedimentary Blue Star Formation conformably overlies the Lewis Lake Formation, and is in turn conformably overlain by the mixed volcanic/sedimentary Bursey Point Formation. The southernmost rocks in the area consist of a mixed mafic and subaerially-erupted silicic volcanic unit (Charles Lake sequence; Swinden, 1988) that has an unknown relationship with the New Bay Pond section of the Wild Bight Group. We consider it likely that the Charles Lake sequence is much younger than other rocks of the Wild Bight Group, and may be correlative with the Silurian (Williams, 1967) Lawrenceton Volcanics (Kusky, 1985; Swinden, 1988) or somehow related to the Silurian Hodges Hill/Mount Peyton Plutons.

Lewis Lake Formation. The Lewis Lake Formation outcrops in a belt around the western part of Lewis Lakes and south of the northern part of New Bay Pond (Fig. 4). Its type section (Kusky, 1985) is located along the river between Lewis Lake and Frozen Ocean Lake. The base of the Lewis Lake Formation is marked everywhere by a fault, succeeded immediately by a heterogeneously deformed unit of fine-grained recrystallized mafic lava, locally containing 1–20 cm randomly-oriented blocks of white-weathering lava. Southwestward-younging mafic volcanic flows, breccias and pillows are the most abundant rock types in the Lewis Lake Formation (Fig. 6a). Flows are typically massive in appearance, and exhibit a brown weathering color. Pillows are less-abundant than massive flows, although an anastomosing cleavage locally developed in the massive flows mimics pillow shapes in outcrop. Mafic volcanic breccias are assigned a hyaloclastite origin (cf. Parsons, 1968; *sensu* Rittmann, 1958) because some of the angular fragments are clearly identifiable as pillow fragments.

Silicic volcanic deposits are locally interdigitated with mafic lavas of the Lewis Lake Formation. White-weathering, black to dark green rhyolites containing large prismatic quartz phenocrysts form isolated domes at several places in the map area. Large lobes, tongues, flow banding, brecciated flow tops and bases, and pillows apparent in many flows, along with thick flow-laminated skins on individual lobes, and abundant large vesicles, suggest subaquatic extrusion of these silicic flows (Dimroth *et al.*, 1979; De Rosen-Spence *et al.*, 1980).

Mafic volcanoclastic and pyroclastic rocks, interbedded with homogeneous mafic flows and pillows, compose a minor component of the Lewis Lake Formation. A distinctive blue-green pyroclastic framework-supported sandstone containing abundant mafic volcanic clasts and

fragmental plagioclase grains, is locally interbedded with the mafic lavas. Rare vesicular basalt lapilli, euhedral plagioclase crystals, and the apparent agglutination of some clasts suggests a pyroclastic, rather than a sedimentary, origin for this unit.

Blue Star Formation. The Blue Star Formation is a thin, entirely sedimentary unit, which coarsens upwards and towards the northwest (Fig. 4). Its type area (Kusky, 1985) is located along the western shore of New Bay Pond. Basal members of the Blue Star Formation, consisting of blue, green, red and maroon cherts were deposited conformably on top of basalts of the Lewis Lake Formation. The cherts are overlain by coarse- to fine-grained mafic volcanoclastic rocks rich in chert pebbles (Fig. 6b). Some pebbles are larger than 10 cm near the top and western ends of the outcrop area, and some of the chert-clast layers exhibit features of syn-sedimentary deformation.

A lesser proportion of the Blue Star Formation consists of resedimented tuffaceous rocks, including a gray-green siltstone containing flattened black flammé. Parallel-laminated purple and green cherty rocks, possibly fine-grained silicic tuffs or bentonites, are dissected by several channels containing coarser grained volcanoclastic deposits. Paleocurrent indicators from the Blue Star Formation indicate a source that lay to the northwest (Kusky, 1985), consistent with the general coarsening and thickening of the formation in this direction.

Bursey Point Formation. The Bursey Point Formation is a thick and poorly exposed unit (Fig. 4), consisting of mixed volcanic flows, tuffs, and volcanoclastic rocks (Fig. 6c). Its type

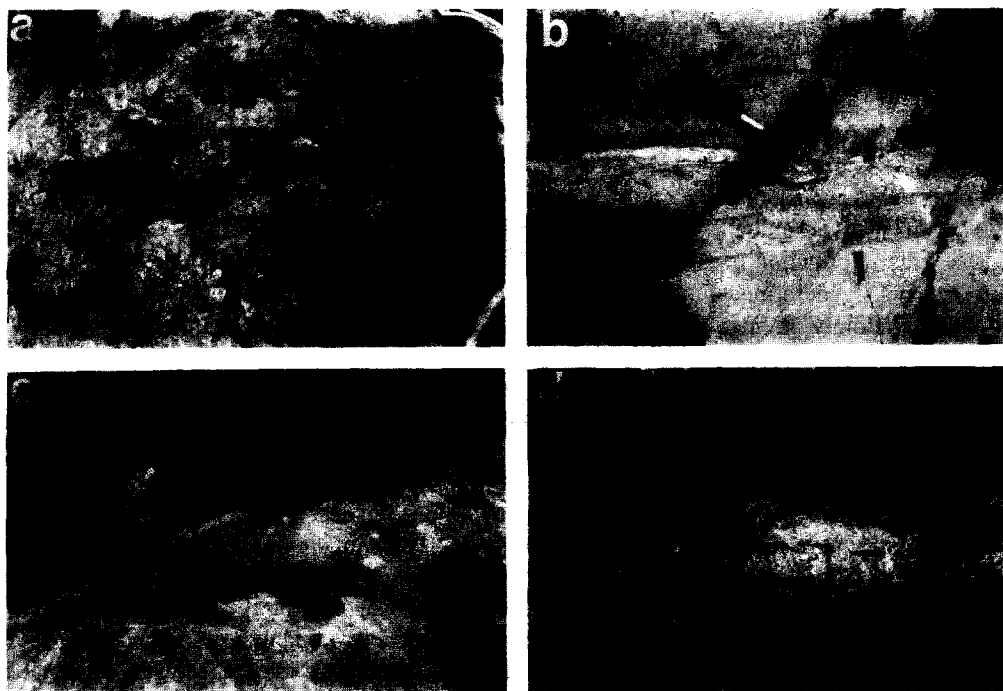


Fig. 6. Characteristic rock-types of the New Bay Pond section of the Wild Bight Group: (a) shows pillow lavas of the Lewis Lake Formation; (b) shows volcanoclastic sedimentary rocks of the Blue Star Formation; (c) shows lithic tuffs of the Bursey Point Formation; and (d) shows felsic ignimbrites of the Charles Lake Sequence.

area (Kusky, 1985) is located along the west shore of New Bay Pond. The base of the Bursey Point Formation is defined as the base of the first volcanic flow overlying volcanoclastic sedimentary rocks of the Blue Star Formation. Each of Frozen Ocean Lake this contact is marked by a massive silicic volcanic flow, with numerous gas escape tubes defining the paleovertical. These features are interpreted to have formed shortly after eruption, during vapor-phase mineralization (Fisher and Schmincke, 1984). An unusual and distinctive "isolated pillow breccia" (*sensu* Carlisle, 1963) occurs near the base of the Bursey Point Formation, and is characterized by small (1 mm–5 cm) basaltic lava stringers and isolated pillow fragments enclosed in a chaotic matrix. The formation of pillows with similar morphology has been attributed to leakage from larger pillows (Moore, 1975) and submarine lava fountaining (Carlisle, 1963; Fisher and Schmincke, 1984). The top of the Bursey Point Formation is not exposed, and it may be intruded by the Hodges Hill/Twin Lakes Complex.

Silurian (?) subaerial volcanic rocks of the Charles Lake sequence. Rocks of the Charles Lake sequence (Swinden, 1988) consist of a suite of bimodal subaerial felsic and mafic volcanics. The type area (Kusky, 1985) for the Charles Lake sequence is located west of the old lumber road along the west shore of New Bay Pond (Fig. 4). Neither the base nor the top of the formation is exposed. Rock types include red, white and orange weathering ash-flow and crystal tuffs, massive rhyolites, ignimbrites, quartz-feldspar porphyries, agglomerates, and massive to pillowed mafic flows (Fig. 6d). The predominantly intermediate to felsic compositions of the Charles Lake sequence differ markedly from the rest of the Wild Bight Group, suggesting that they are unrelated.

Wild Bight Group (Northeastern section)

The northeastern-most rocks in the New Bay Pond area are equated with the Seal Bay Head Formation (Nelson, 1979) of the Wild Bight Group (Figs 4 and 5). Included in the type section of this formation at Seal Bay Head on the mouth of Seal Bay (Fig. 2) are mafic pillow lavas, volcanic breccias, green volcanoclastic sandstones and argillites (Nelson, 1979). In the map area, the Wild Bight Group consists of green and brown volcanoclastic sandstone, some tuffaceous units (with structurally flattened pumice fragments), manganiferous siltstone, conglomerate, basalt, mafic volcanic breccia, and tan, blue and white weathering cherts, and dark blue to black argillites. The top of the Wild Bight Group passes conformably upwards into red cherts of the Shoal Arm Group on an island in Moose Cove at the north end of New Bay Pond.

Shoal Arm Formation. Cherts and argillites of the Shoal Arm Formation (Espenshade, 1937) are repeated by folds and faults throughout the Notre Dame Bay area (Sampson, 1923). Graptolites from the Shoal Arm Formation, and its correlatives throughout the Exploits Subzone, yield ages from Llandeiliian–Caradocian, and locally Ashgillian (Helwig, 1969; Dean, 1978; Kusky, 1985; Williams, 1991; Brüchert *et al.*, 1994). Espenshade's (1937) original definition of the Shoal Arm Formation included an upper clastic unit which is now regarded as a separate formation known as the Sansom Graywacke (and equivalents). The thickest section of the Shoal Arm Formation occurs in the New Bay Pond area (Dean, 1978) and a more detailed stratigraphic subdivision of these rocks is presented here. We informally subdivide the Shoal Arm Formation into three members, including from base to top, the Moose Cove member, White Point member, and the Drowning Point member. The type area for these three members is designated at the north end of New Bay Pond (Figs 4 and 7). These three members are recognizable and mappable throughout the Notre Dame Bay area, and because of their

distinctive character and abundance of graptolites, they provide a regional stratigraphic marker horizon.

The Shoal Arm Formation is conformably underlain by volcanic and volcanoclastic rocks of the Wild Bight Group, and although its upper contact in the New Bay Pond area is tectonic, workers elsewhere in the region have reported a gradational change into the overlying Point Leamington Formation (Brüchert *et al.*, 1994). Thus, the regional stratigraphy has been preserved across this contact, even though it is the site of localized bedding-parallel slip. Soft-sediment deformation is a nearly ubiquitous feature of rocks of the Shoal Arm Formation. Here, we follow the suggestion of Dean (1978) and Nelson (1979) that the top of the Shoal Arm Formation be defined as the base of the first graywacke bed overlying the black shales.

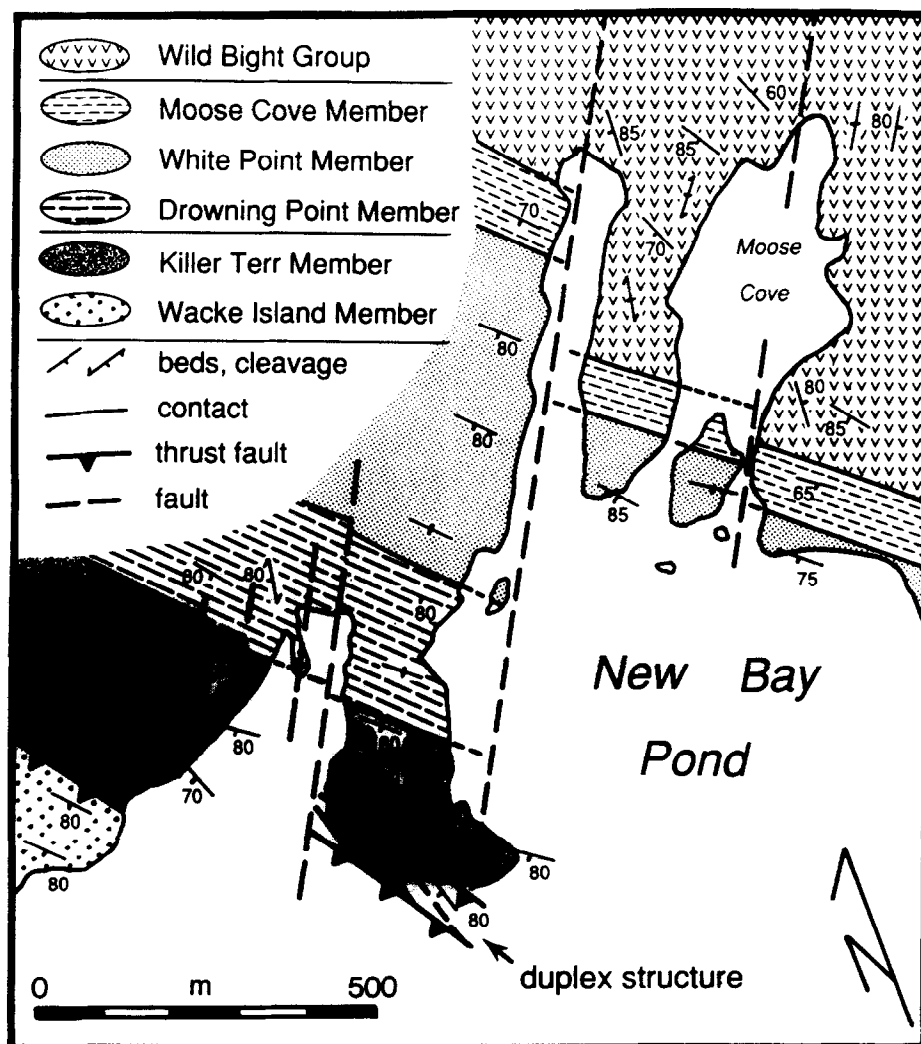


Fig. 7. Map of the northeastern part of New Bay Pond, showing late dextra strike-slip faults offsetting Shoal Arm Formation stratigraphy, and duplex structure on Drowning Point. Location of map indicated by box in Fig. 4.



Fig. 8. Cherts and argillites of the Shoal Arm Formation: (a) shows interbedded white cherts and argillites of the White Point member; and (b) shows red radiolarian cherts of the Moose Cove member.

Moose Cove Member

The Moose Cove member is the most distinctive part of the Shoal Arm Formation, consisting of bright red radiolarian cherts containing carbonate and manganese nodules (Fig. 8a). Bed thicknesses range from <1 cm to >10 cm, and parallel laminations are ubiquitous. Thin (<5 cm) iron oxide stained clastic horizons with Bouma sequences are present, and these contain small (<3 mm) fragments of altered basalt and volcanic glass, broken and altered feldspar grains (with albite twins), quartz and pyrite. These clastic beds are poorly sorted and grain supported, with grains ranging from angular to well-rounded. The matrix is composed of microcrystalline quartz grains identical to the surrounding chert beds. The top of the Moose Cove member grades over a distance of 5–10 m into the White Point member (Fig. 6), with rocks of both units interbedded over this interval. Some lenticular chert layers with a red tinge are also present in this transition zone.

White Point Member

The White Point member contains white-weathering blue-gray bioturbated chert interbedded with argillite layers, which are more abundant near the top of the member. Steely blue manganese and rusty-orange iron weathering horizons are common. Two scales and types of lamination occur in the White Point member: (1) an alternation between chert and argillite; and (2) banding within the chert (Fig. 8b). Banding in chert sequences has been attributed to many factors, including diagenetic alteration, fluctuating productivity of siliceous tests, sedimentary differentiation by turbidity currents, and double accumulation of muddy turbidities and slow siliceous tests, or siliceous turbidites and slow hemipelagic clay accumulation (Davis, 1918; Granau, 1965; Murray, 1992). Since Bouma sequences occur in a few of the chert beds, and the argillites are essentially structureless, a combination of sedimentary differentiation and double deposition appears to have formed the layering in the White Point member. Differentiation by separate turbidity currents is suggested to be responsible for the banding within the cherts,

whereas the argillites are suggested to be hemipelagic oozes. The top of the White Point member grades over approximately 2 m into the overlying Drowning Point member.

Drowning Point Member

The Drowning Point member contains interbedded black shales, argillites, and very rare thin graywacke beds. Individual beds range from sooty dark-black argillites to light-gray, silica-rich shales. Weathered outcrops typically exhibit a yellow to orange weathering stain, related to abundant pyrite. The magnetic properties of the Drowning Point member and the Moose Cove member were used with aeromagnetic maps (proprietary data of Noranda Exploration Co.) to map the distribution of these rock types through areas of sparse outcrop, and to determine their separation across the numerous dextral faults that offset these beds.

Graptolites collected by us from the north end of New Bay Pond (location a, Fig. 4) have yielded the following specimens (identified by J. Riva, pers. comm., 1983): *Orthograptus quadrimucronatus* (Hall), *Leptograptus cf. flaccidus* (Hall), *Dicellograptus sp.*, and *Orthograptus sp.*, typical of upper Caradocian–lower Ashgill (*Dicranograptus cligani*, *Pleurograptus linearis*) zones, or younger fauna from Newfoundland, and is identical to fauna taken from the Moffat area, Scotland (J. Riva, pers. comm., 1983). These fauna indicate an age for the upper Shoal Arm Formation that is significantly younger than that suggested by Dean (1978), on the basis of the following graptolites extracted from the same outcrop by H. Williams, and identified by L. M. Cumming (Dean, 1978): *Leptograptus flaccidus* (Hall), *Climacograptus sp.* either *bicornis* (Hall) or *diplacanthus* (Bulman), *Orthograptus sp.* The compiled ages are in agreement with the Upper Caradoc–Lower Ashgill graptolite fauna reported by Williams (1991) from the Point Leamington Formation in the northern Exploits Subzone.

Point Leamington Formation (Sansom correlative). The Point Leamington Formation is an upward-coarsening flysch-type sedimentary sequence with correlatives throughout the Exploits Subzone, known generally as Sansom-equivalent strata (Nelson, 1979). In the New Bay Pond area, the Point Leamington Formation occurs as a northwesterly tapering wedge bounded below by the Shoal Arm Foundation and above by a fault surface that progressively cuts out portions of the stratigraphy toward the northwest. Here, the Point Leamington Formation is informally divided into two members: the basal Killer Terr member, composed of alternating shale and graywacke beds, and the Upper Wacke Island member, containing massively bedded quartz-rich volcanic-lithic wackes, and only minor shale beds. The type sections for these two members are continuous with the type sections for division of the Shoal Arm Formation, at the north end of New Bay Pond (Fig. 4).

Killer Terr Member

Rocks composing the Killer Terr member include buff-weathering graywackes interbedded with more abundant black shales. Where measurable, the ratio of sand:shale ranges from 1:10 to 3:1. Higher ratios occur near the top of the member, suggesting a progressively increasing supply of clastic detritus during deposition of the lower Point Leamington Formation. Although the Killer Terr member has suffered a unique and complex deformation history, the regionally recognized stratigraphy appears to have been preserved across this unit.

The contact between the Point Leamington and Shoal Arm Formations is structurally complex and where the outcrop is abundant enough to permit detailed mapping, this contact is marked by a mélangé developed in the Killer Terr member. The area best-exhibiting this deformation is

exposed at the northeastern end of New Bay Pond where the structural architecture is that of a duplex structure (Fig. 7). From the Shoal Arm Formation below, deformation gradually increases into and through a lower, thinly-bedded argillaceous member of the Point Leamington Formation (Kusky, 1985). A fault contact is recognized between this lower member and an upper, massively bedded graywacke member of the Point Leamington Formation.

Highly variable styles of deformation and orientations of structures are recognized within the Killer Terr member (Fig. 9). Dismembered isoclinal folds, and folds with large variations in hinge line orientation are common, but an overall southward vergence is present. Detached blocks of graywacke are isolated in an argillaceous matrix, and many graywacke beds are extended and boudinaged, forming a Type 1 *mélange* (*sensu* Cowan, 1985). The upper Caradocian–lower Ashgillian age of graptolite fauna (see above) from this unit places a lower limit on the age of deformation in the Killer Terr member.

Wacke Island Member

Sedimentary rocks of the Wacke Island member include massive light brown to buff weathering sandstones, which are distinctive from other graywackes within the Wild Bight and New Bay Groups (and correlatives) because of the abundance of detrital quartz grains and silicic volcanic pebbles which are rare or absent in pre-Caradocian sedimentary rocks (Helwig, 1967, 1969; Nelson, 1979). The graywackes are typically massively bedded, with individual bed thicknesses ranging from <1 cm to >3 m. Paleocurrent indicators including cross-laminations and flute marks, are fairly abundant (Kusky, 1985) and suggest transport from a source terrane that lay to the north or northwest (present coordinates). Structures indicative of paleoslope (slumps and flame structures, Fig. 10) are also abundant, and suggest that the basin sloped towards the southeast (Kusky, 1985).

The graywackes include both matrix- and grain-supported varieties, with a large variation in the amount of matrix material present. Detrital grains include abundant quartz, some with hexagonal prismatic morphology, demonstrating a volcanic origin. Silicic magmatic lithic

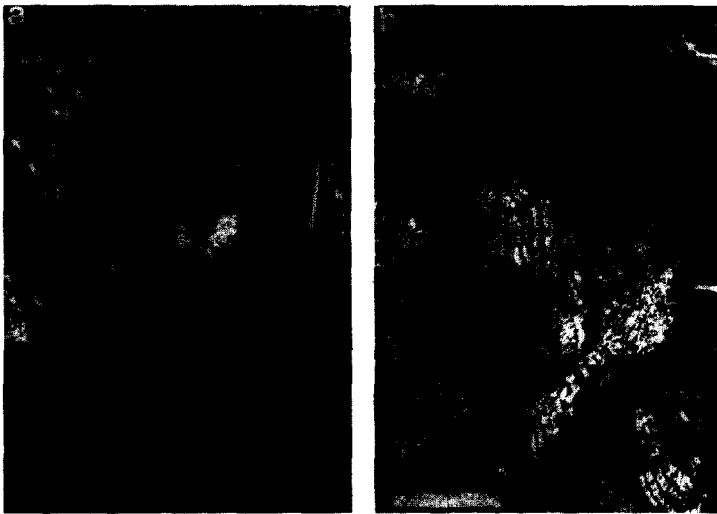


Fig. 9. (a) Rootless isoclinal fold in Killer Terr Complex; and (b) Folds above a décollement in the Killer Terr Complex (scale bar represents 50 cm).

fragments are also abundant, some with large grain sizes indicating erosion from a shallow plutonic source. Rare mafic volcanic clasts occur, as do coarse-grained polycrystalline quartz aggregates, interpreted as vein quartz. Opaque grains are common, and include rare chromite (Kusky, 1985), suggesting that in the source area an ophiolitic terrane was eroded (Nelson and Casey, 1979).

STRUCTURAL ANALYSIS

Rocks of the Frozen Ocean Lake, New Bay Pond area comprise a southwestward-younging sequence exposed on the southwest limb of the Seal Bay fold, a northwest trending anticline offset by numerous, dominantly right-lateral strike-slip faults (Fig. 2). Strata strike west-northwest and are dominantly vertical to steeply southwestward dipping (Fig. 11). The

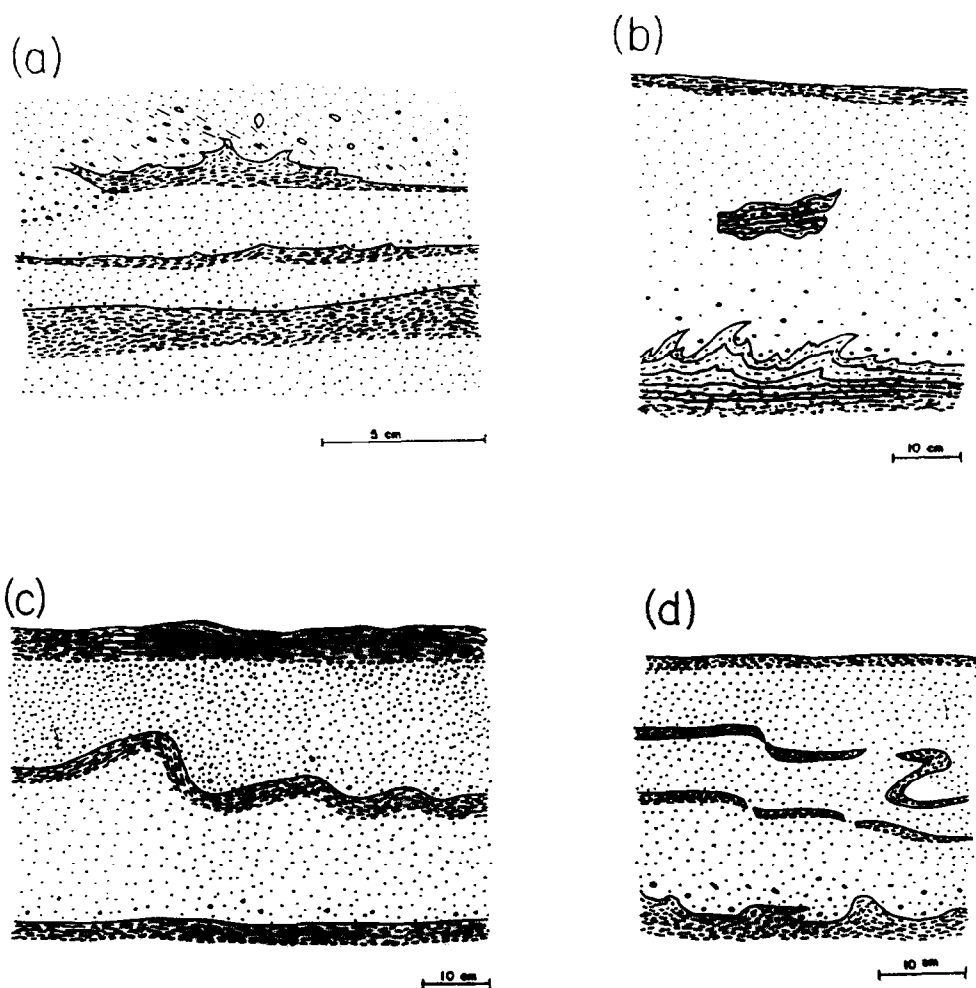


Fig. 10. Soft-sediment deformation features in the Point Leamington graywacke including (a) flame structures and erosive channels, forming proto rip-up clast; (b) rip up clast; (c) shale rip-up clasts in layer over flame structures; and (d) asymmetric folds bounded by planar layers, interpreted as soft-sediment structures.

northwest strike may be related to folding about a vertical axis, perhaps induced by motion on bounding dextral strike-slip faults (Kusky, 1985; Bostock, 1988; van der Voo *et al.*, 1991). Foliation orientations within the New Bay Pond section of the WBG and structurally underlying sedimentary rocks show two concentrations, particularly evident in the WBG (Fig. 12). These point maxima correspond to (1) the regional northeast-striking spaced to dominal cleavage; and (2) a northwest-striking anastomosing to sinuous foliation that is spatially restricted to areas near the Fog Fault. A rare northwest-striking spaced cleavage is parallel to the axial trace of the Seal Bay fold.

Rocks of the Frozen Ocean Lake, New Bay Pond area are metamorphosed to prehnite-pumpellyite and greenschist facies (Kusky, 1985), and primary structures are readily discernible. Cross-cutting relationships between structures and intrusive rocks of the Hodges

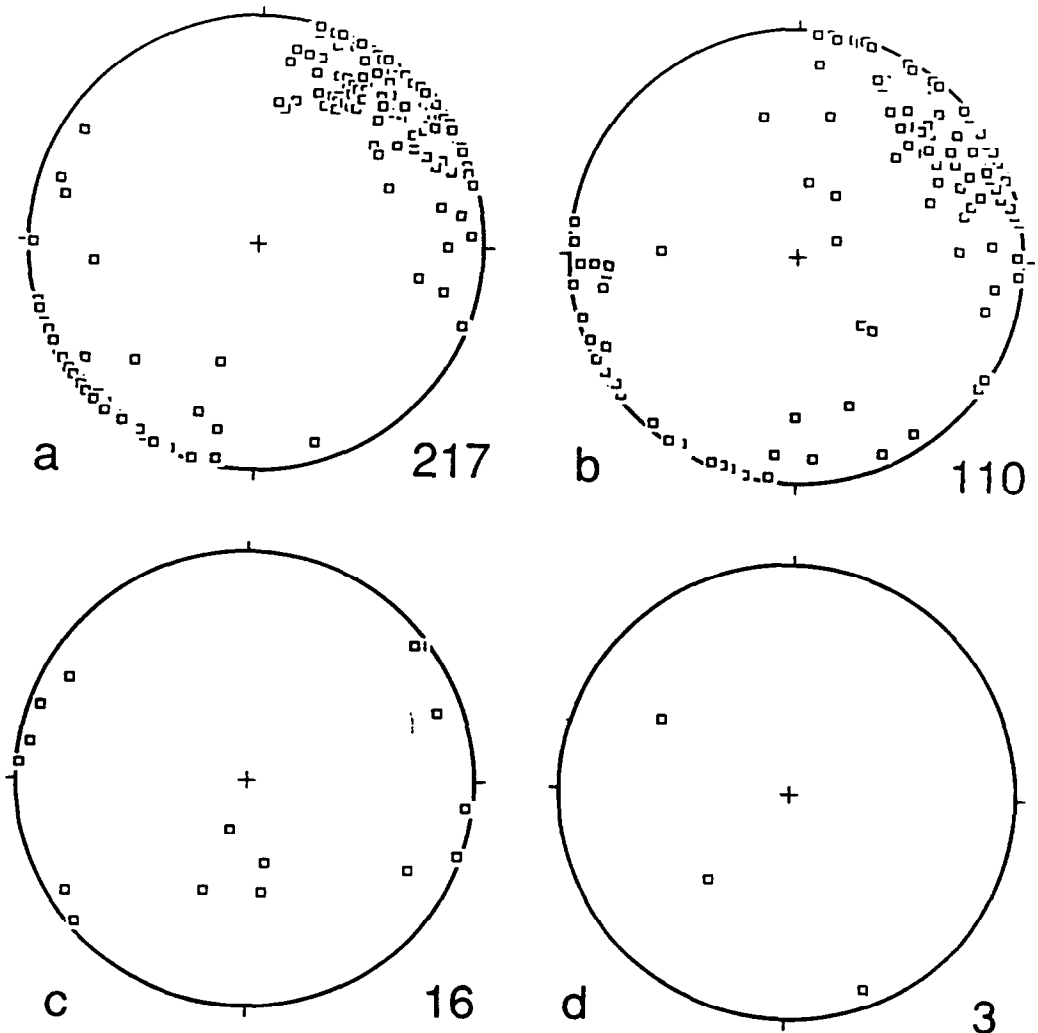


Fig. 11. Lower hemisphere equal area projections of bedding orientations from (a) the Point Leamington and Shoal Arm Group sedimentary rocks; (b) the New Bay Pond section of the Wild Bight Group; (c) the northern section of the Wild Bight Group; and (d) the Charles Lake sequence (*sensu* Swinden, 1988).

Hill batholith enable us to divide the structures into two main groups: those that predate the probable Late Silurian–Early Devonian batholith, and those that post-date the batholith.

Ordovician and Early Silurian structural development

Early soft-sediment structures. Many of the zones of stratal disruption in the New Bay Pond area have characteristics indicative of deformation while the sediments were soft (cf. Maltman, 1984 and Pickering, 1987). Soft-sediment folds are especially common in the Shoal Arm and Point Leamington Formations in the New Bay Pond area. Figure 10 shows several examples of syn-sedimentary and soft-sediment structures sketched from outcrop photographs, all consistently indicating south or southeast facing slopes and down-slope current flow. Overturning directions and asymmetries of most slump folds are consistent with basal slopes toward the

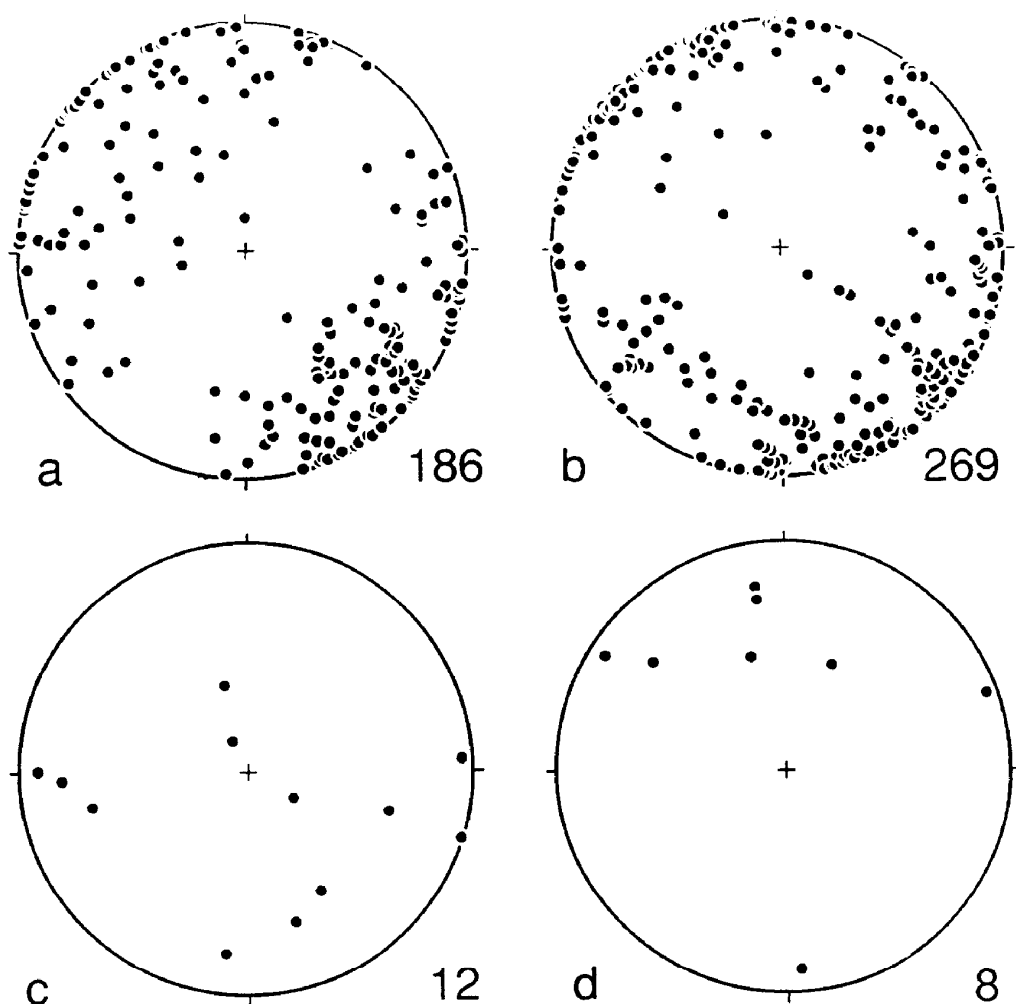


Fig. 12. Lower hemisphere equal area projections of poles to cleavage planes in (a) the Point Leamington and Shoal Arm Foundation sedimentary rocks; (b) the New Bay Pond section of the Wild Bight Group; (c) the northern section of the Wild Bight Group; and (d) the Charles Lake sequence (*sensu* Swinden, 1988).

south or southeast (Kusky, 1985). Similar south- to southeast-directed paleocurrent and paleoslope indicators have been reported from elsewhere in the Sansom-equivalent strata (Helwig, 1967, 1969; Helwig and Sarpi, 1969; Nelson, 1979; Arnott, 1983; Arnott *et al.*, 1985; Karlstrom *et al.*, 1983b; Pickering, 1987; van der Pluijm, 1986).

Late Ordovician/Early Silurian thrust faults and folds

The earliest demonstrably tectonic structures in the Frozen Ocean Lake area include southward directed thrust faults with associated Late Ordovician-Early Silurian clastic wedges, regional folds about an approximate east-west axis, and related small-scale structures including abundant isoclinal intrafolial folds. The contact between rocks of the Wild Bight Group (Lewis Lake Formation of Kusky, 1985) and structurally underlying sedimentary rocks is marked everywhere by an early fault (the Fog Fault; Kusky and Kidd, 1985) which is well-exposed in at least 8 locations in the map area (Kusky, 1985). In many places, the Fog Fault is marked only by a lineament associated with fault breccia, adjacent to the fault footwall, beds are overturned and young away from the fault trace (e.g. location 1, Fig. 4). The fault exhibits both brittle and ductile features, either of which may dominate the outcrop character at any given location.

Excellent exposures of the Fog Fault crop out in small coves south of the northern portion of New Bay Pond (location 2, Fig. 4), where exposure is continuous from the Point Leamington Formation through the Fog Fault and associated fold, into mafic volcanics of the Wild Bight Group. At locality 2, the Fog Fault is cut by a 10 cm wide undeformed mafic dike presumably associated with the Hodges Hill/Twin Lakes batholith complex, possibly constraining the timing of movement as pre-Late Silurian. The fault zone is about 5 m wide (Fig. 13a) and is composed of strongly foliated fault breccia and cross-cutting cataclastite and pseudotachylite indicating that in this location the Fog Fault has dominantly brittle characteristics. Foliation gradually decreases in intensity away from the fault, whilst it is most intense in the hanging wall. Rocks of the Lewis Lake Formation on the southern side of the fault are, for several meters away from the fault, brecciated and cut by thin cataclastic gouge zones.

Fault zone parallel and oblique foliations (Fig. 13b), interpreted as S-C banding (*sensu* Berthe *et al.*, 1979; Lister and Snoke, 1984; Hanmer and Passchier, 1991), and field-scale asymmetric pressure shadows indicate south-side-down relative sense of shear. Cataclastic fault breccia zones preserve an imbricate-type of composite planar fabric (Fig. 14a), with pseudotachylite and through-going shear surfaces oriented parallel to the shear zone boundaries, and with the fragment imbrication oriented with the same obliquity as the S-C fabric in nearby ductile domains of the fault zone. Movement is assumed to have occurred parallel to a mineral rodding lineation developed in the fault rocks (Bryant and Reed, 1969) that plunges 45° southeast. The lineation is approximately parallel to hinge orientations of NW striking isoclinal folds from locality 1, suggesting that folds associated with this fault have been rotated into the movement direction along the fault plane, and that they are no longer parallel to their cut off lines (Skjærna, 1979; Ramsay and Huber, 1987).

Fault locality 3 along an abandoned lumber road at the northwestern end of New Bay Pond (Fig. 4) also offers nearly continuous exposure across the fault and exposes chloritic mafic volcanics of the Lewis Lake Formation, altered, rusty-weathered silicified fault gouge, and graywackes of the Point Leamington Formation that young away from the fault zone. The fault rocks exhibit a pervasive vertical foliation which strikes parallel to a pronounced lineament in this area that separates the Lewis Lake Formation from the Point Leamington Formation. Microscopic examination reveals that the fault rocks are composed of silicified fault gouge

containing abundant pyrite and carbonate veins. No sense of shear indicators were obtainable from this locality.

Locality 4 on the southwestern side of Lewis Lake (Fig. 4) exhibits a well-defined northwest striking topographic lineament separating Wild Bight Group rocks from the Point Leamington Formation. Although outcrop in this valley is sparse, fault breccia boulders are common in the stream bed, and rocks on either side of the lineament are more strongly foliated than is typical of either rock unit. The Fog Fault is exposed at locality 4 where it dips 55° west and is composed of rusty-weathered silicified fault gouge similar to that at locality 3. Microscopic asymmetric foliations and pressure shadows around porphyroclasts (Fig. 14b) and S-C composite planar fabrics suggest a west side down displacement sense, parallel to the stretching lineation.

At locality 5 (Fig. 4) the spacing and intensity of a northeast-striking vertically-dipping cleavage increases in intensity adjacent to the fault, and cleavage planes are locally coated with southwestward plunging slickenlines. Immediately adjacent to the trace of this fault, Point Leamington Formation strata young away from the contact with the Lewis Lake Formation.



Fig. 13. (a) The Fog Fault at location 3 (Fig. 4), showing large porphyroclasts in a dark matrix; (b) the Fog Fault showing weathered S-C composite planar fabric.

Beds are folded in a sheath-like manner, plunging approximately 30° southwest. The Fog Fault outcrops 50 m inland from the shore, where it cuts stratigraphically up through the Point Leamington Foundation toward the south, while beds of the Lewis Lake Formation remain parallel to the curved trace of the fault. This geometry is interpreted as a hanging wall flat over a footwall ramp that was subsequently rotated during regional folding (to form the Seal Bay fold) to expose a cross-sectional view (e.g. Bradley, 1989). The ramping-up to the southeast could explain the apparent thickening of the Point Leamington Formation in this direction.

At locality 6 on Lewis Lake (Fig. 4), deformed mafic volcanic rocks of the Lewis Lake Formation adjacent to the fault zone are well exposed. The Point Leamington Formation again youngens away from the fault for a distance of approximately 10 m, where an isoclinal fold hinge is exposed, with axial plane parallel to the fault trace, locally offset by kink folds. Within the fault rocks, a fault-parallel foliation surrounds large porphyroclasts with asymmetric pressure shadows (Fig. 14b) suggesting that the Wild Bight Group moved down and toward the southeast with respect to the Point Leamington Formation.

Pre-regional folding attitude of the Fog Fault. All stretching lineations associated with the Fog Fault are plotted on a lower hemisphere equal area projection (Fig. 15a) and restored to their



Fig. 14. (a) Fault breccia intruded by pseudotachylite (sample is 12 cm across); and (b) asymmetric pressure shadows around a porphyroclast from the Fog Fault.

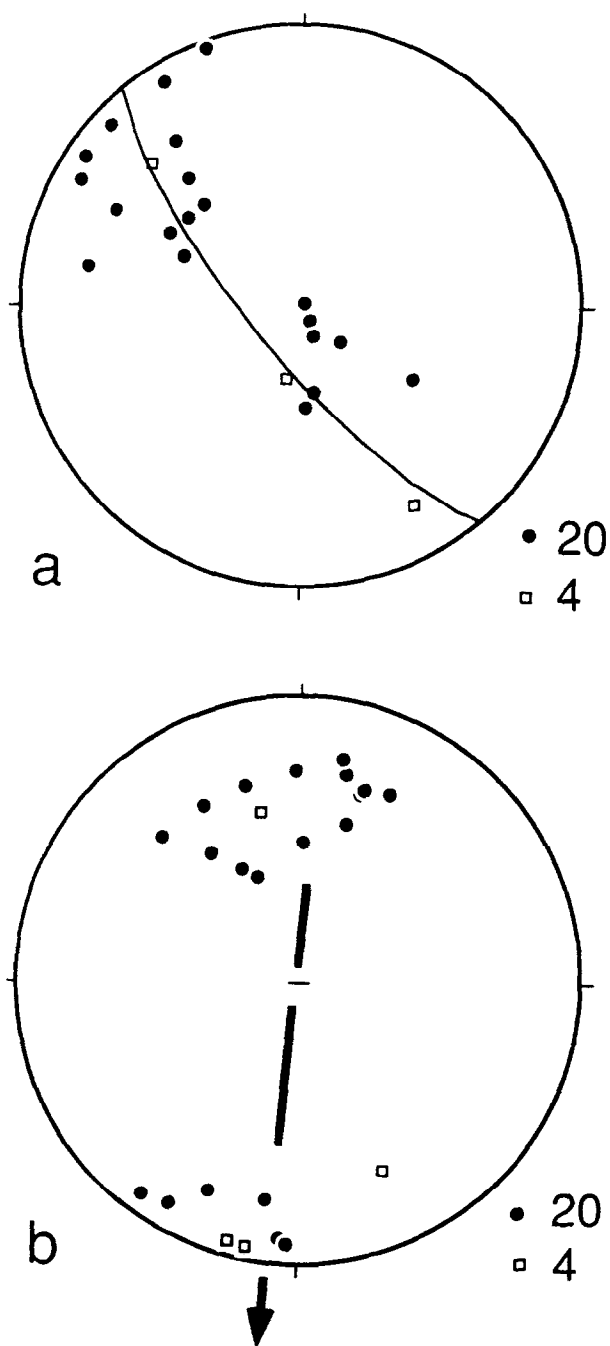


Fig. 15. (a) Lower hemisphere equal net showing orientation of linear elements associated with the Fog Fault (solid circles=lineations, squares=fold hinges); (b) lower hemisphere equal area net showing D1 lineations from the Fog Fault, after bedding is rotated back to horizontal. Arrow shows inferred sense of thrusting of hanging wall over footwall.

presumed initial orientation by a simple rotation of bedding back to horizontal using the strike of the Seal Bay antiform (Fig. 2). This effectively removes tilting imposed by later folding, although it does not take into account any deviation of bedding planes from the horizontal incurred during early deformation, rotations associated with transpression, nor any deviations from a flexural slip model of folding. The result (Fig. 15b) shows two point maxima indicating nearly horizontal north-south lineations. The group that plunges shallowly to the north has a slightly steeper plunge than the group that plunges to the south; when considered with the map pattern, which shows several footwall ramps cutting stratigraphically up toward the southeast, it is apparent that the horizontal lineations are associated with thrust flats, and the gentle N-plunging lineations are associated with thrust ramps. Together with the sense of shear indicators, this geometry implies that the New Bay Pond section of the Wild Bight Group was thrust from north to south over the Point Leamington Formation (Fig. 15b). If the regional flexure in Notre Dame Bay (Fig. 2) is straightened out parallel to the overall strike of the orogen (assuming it is a deformational rather than a primary feature), then this thrusting direction is changed from southward to southeastward.

Deformation of the Shoal Arm Formation/Point Leamington Formation contact may also be related to this early thrusting event. Orientations of fold hinge lines, slickenline fibers, and fault planes within the Killer Terr Complex show considerable scatter (Fig. 16a), even when restored by rotation of bedding planes back to horizontal. However, when planar and linear structures associated with late, cross-cutting faults are removed (Fig. 16b), a more consistent picture of early deformation geometry emerges (Fig. 16c). One set of faults was originally subhorizontal, whereas the second set of faults, within the duplex structure, was originally northwest dipping. These faults are cut by dikes related to the Hodges Hill pluton, further constraining the timing of movement as pre-Late Silurian–Early Devonian. Slickenlines indicate a south-southeast sense of movement on essentially horizontal planes, and fold hinge lines, whether they had a slump or tectonic origin, are partially rotated into the movement direction along the faults. Most of the structures at the base of the Point Leamington Formation are thus suggested to be related to an early deformation episode involving south or southeast directed thrusting along the Killer Terr Complex. The near-parallelism between the sense of movement that produced these structures, and the movement direction on the Fog Fault, suggests a common origin.

REGIONAL EXTENT AND TIMING OF SOUTHEASTWARD DIRECTED THRUSTING

There is considerable evidence that the thrusting event that placed rocks of the Wild Bight Group over Caradocian sedimentary rocks in the New Bay Pond area is part of a regional Late Ordovician/Early Silurian tectonic episode. Helwig (1967) recognized a pre-Devonian generation of folds about north to northwest trending axes in the New Bay area (Fig. 2), and suggested that several of the important faults in that area formed during this (his F_2) deformation episode. Nelson (1979, 1981) described NNW dipping syn-sedimentary thrust faults from the Badger Bay/Seal Bay area (Fig. 2), and attributed them to a post-Middle Ordovician and pre-Devonian back-thrusting event. Dean (1978) did not recognize any pre-Devonian structures east of the Halls Bay Fault (Fig. 2) but Dean and Strong (1977) suggested that the Lobster Cove and Chanceport faults in Notre Dame Bay may have formed during the deformation event here attributed to Early Silurian thrusting. Karlstrom (1983) and Karlstrom *et al.* (1982, 1983a) recognized widespread F_1 related isoclinal and intrafolial folding associated with thrusting, and proposed that the entire eastern Notre Dame Bay area is allochthonous. Arnott (1983) described NNW-dipping syn-sedimentary thrust faults, and Pickering (1987) described similar structures

including “wet-sediment” faults related to these events. Post-Middle Ordovician thrust faults were also reported from the Exploits Subzone by Kay and Williams (1963), Jacobi and Schweickert (1976), Kidd (1977), Blackwood (1979, 1982), Pajari *et al.* (1979), Colman-Sadd (1980), Thurlow (1981), Colman-Sadd and Swinden (1984), van der Pluijm, (1984, 1986, 1987), van der Pluijm and van Staal (1988), Kusky (1985), Kusky and Kidd (1985), Elliott (1985a); and Elliott and Williams (1988). Most workers have reported south- or southeast-directed thrusting

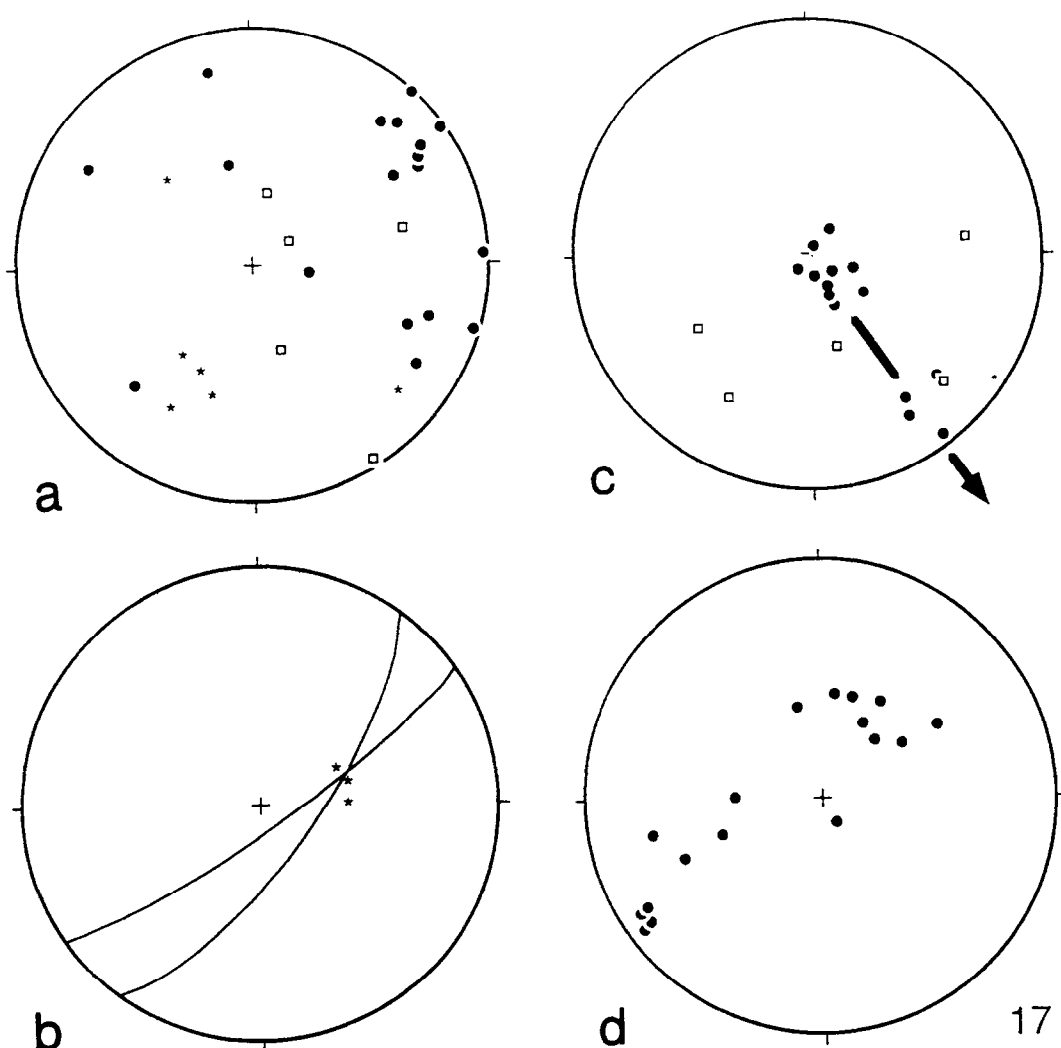


Fig. 16. (a) Lower hemisphere equal area net showing the orientation of poles to faults (solid circles), slickenlines (asterisks), and fold hinges (squares) in the Killer Terr Complex; (b) shows the orientation of younger structures which cross-cut the duplex; (c) shows orientation of early structures with the bedding restored to the horizontal using a simple strike rotation. Two sets of faults are associated with the duplex structure at the base of the Point Leamington Formation; the first was subhorizontal and parallel to bedding plane, whereas the second dipped shallowly northwest, suggesting south-southeast directed thrusting (as indicated by arrow). Fold hinges show partial rotation into shearing direction; (d) lower hemisphere equal area net showing orientations of slickenlines associated with late fault zone.

(Nelson, 1981; Thurlow, 1981; Blackwood, 1982; Colman-Sadd and Swinden, 1984, Kusky and Kidd, 1985; Lafrance and Williams, 1992), but Karlstrom *et al.* (1982, 1983a) suggested northwestward directed thrusting of the Botwood Group from southeast of the New World Island area (Fig. 2). Van der Pluijm (1986, 1987) and Elliott (1985b) proposed to resolve this discrepancy by invoking two different thrusting events, one which occurred before, and the other after deposition of the subaerial Botwood Group.

The ages of the initiation and cessation of thrusting in the Exploits Subzone are constrained by the age of thrust-related clastic wedges and U-Pb dates on cross-cutting plutonic units. Sansom-equivalent strata including the Point Leamington Formation range in age from upper Caradocian-Ashgillian as determined from graptolites in interbedded slates (Bergstrom *et al.*, 1974; Williams, 1991). Arnott (1983) and Pickering (1987) have shown through paleocurrent analysis that the Sansom equivalent strata were derived from the north or west, consistent with thrusting from that direction. A Llandoveryan age for the Goldson Conglomerate (Twenhofel and Shrock, 1973; Williams, 1962; Helwig, 1967) is based on the preservation of corals, brachiopods, trilobites, and other fossils within limestone boulders in an olistostromal horizon at the base of the conglomerate, and on brachiopods in the slaty matrix to the olistostromes (McKerrow and Cocks, 1977). Elliott *et al.* (1991) report that thrusting began on New World Island in the Llandoveryan or perhaps as early as Ashgillian (Elliott, 1985a), and U-Pb dates on cross-cutting intrusions show that thrusting ended there by 408 ± 2 Ma (Pridolian, or Late Silurian). Cross-cutting plutonic rocks presumably of the Hodges Hill Complex show that thrusting ended in the New Bay Pond area by Late Silurian–Early Devonian. Further south in the Buchans area, thrusts involve $473 \pm 3/-2$ Ma and $462 \pm 4/-2$ Ma rocks (Whalen *et al.*, 1987; Dunning *et al.*, 1987), and a 429 ± 3 Ma pluton cuts rocks of one thrust sheet (Whalen *et al.*, 1987). It is unknown, however, whether or not this pluton is allochthonous. In south-central Newfoundland a late Arenigian deformation event placed ophiolitic rocks over the Gander Zone (Colman-Sadd *et al.*, 1992), although the relationship of this possible Early Ordovician (Penobscottian) deformation event to the Middle Ordovician–Late Silurian events in the northern Exploits Subzone remains unclear.

Williams and O'Brien (1991) reported a middle Llandoveryan (Aeronian) graptolite with European affinities from Sansom equivalent strata in the Bay of Exploits, showing that the Sansom equivalent strata and Goldson conglomerate may be in part lateral equivalents of different facies, and that a large oceanic tract still existed between Laurentia and the Exploits Subzone in Early Silurian times. In summary, there is evidence in the northern Exploits Subzone for regional contraction and thrusting (Fig. 15) lasting from Middle Ordovician (Caradocian) through Lower Silurian (Llandoveryan). This thrusting event contracted the region between the Exploits Subzone and the Gander Zone, accreting material deposited in the open Iapetus Ocean.

SIGNIFICANCE OF REGIONAL SOUTHEAST DIRECTED THRUSTING

Middle Ordovician–Early Silurian thrusting in the Exploits Subzone formed a southeastward-vergent (present coordinates) fold-thrust belt (Arnott *et al.*, 1985; Kusky, 1985; Kusky *et al.*, 1987; Pickering, 1987; van der Pluijm, 1986, 1987; van der Pluijm and van Staal, 1988; Coleman-Sadd *et al.*, 1992; Dunning *et al.*, 1990; Swinden, 1990). The convergence was initially accommodated along a series of thrusts in the Exploits Subzone, including the Toogood Fault, Cobbs Arm Fault, the Dildo Fault, and the Fog Fault (Fig. 17). These faults are now folded, and correlations are hampered by the abundant late dextral faults, but their distribution

throughout the northern Exploits Subzone (Fig. 2) suggests a possible structural correlation of lower portions of the New Bay Pond section of the Wild Bight Group with the South Lake ophiolite (Lorenz and Fountain, 1982), the base of the Tea Arm volcanics, and some of the volcanic sequences on New World Island (Fig. 17). Exploits Subzone thrusting eventually formed a new south-vergent (Early Silurian coordinates) accretionary complex (van der Pluijm,

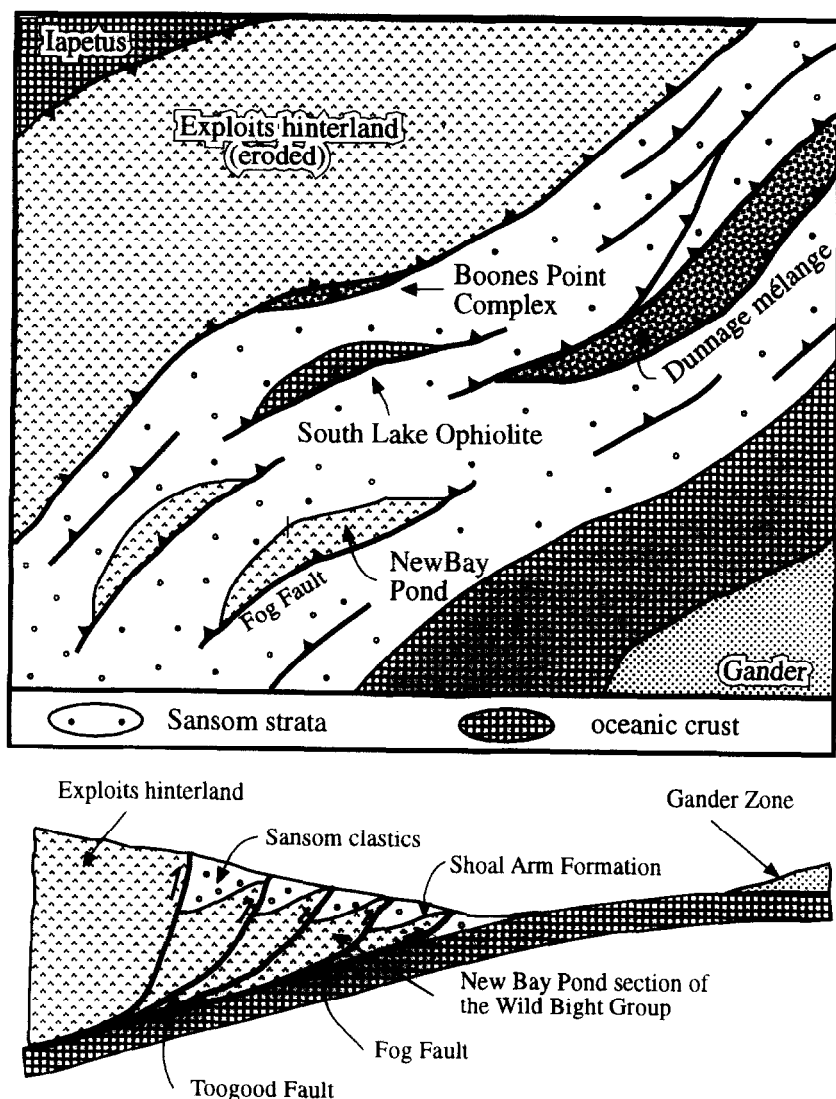


Fig. 17. Schematic Lake Ordovician–Early Silurian paleogeographic reconstruction (a) and cross-section (b) of north-central Newfoundland. Thrusts formed at this time place island arc and locally ophiolitic (South Lake) rocks over similar sequences and a contemporaneous flysch sequence (Sansom graywacke and correlatives, including the Goldson conglomerate). Several examples of these thrusts are shown, including the Fog Fault, Toogood Fault (over the Dunnage mélangé), and the Lukes Arm-Sops Head Fault. Earlier faults (e.g. Lobster Cove, Chanceport Fault and Sops Head, Lukes Arm Fault) are rotated into steeper attitudes than younger faults, as the Exploits Subzone begins being thrust over Gander Zone continental margin rocks. NBP=New Bay Pond.

1986, 1987; van der Pluijm and van Staal, 1998) above a north-dipping (Early Silurian coordinates) subduction zone, placing Exploits Subzone rocks over ophiolitic basement and Gander Zone rocks (Fig. 17).

The stratigraphic sequence preserved in the northern Exploits Subzone (Fig. 3) is that of a classic flexural-load induced subsidence sequence deposited in a foreland basin. Rapid initial subsidence and deepening of the depositional basin is recorded by cherts and argillites of the Shoal Arm Formation, whereas the upward coarsening flysch-type sedimentary sequence of the Sansom-correlative strata and Goldson conglomerate records the approach, uplift, and unroofing of the source terrain in the fold/thrust belt to the north/northwest within the Exploits Subzone.

Middle Ordovician ophiolite obduction associated with the Taconic orogeny in western Newfoundland occurred at circa 485–450 Ma (Stevens, 1970; Fahreaus, 1973; Dallmeyer, 1977; Dallmeyer and Williams, 1975; Dunning and Krogh, 1985). Middle Ordovician–Early Silurian (450–435 Ma) thrust initiation in the Exploits Subzone immediately post-dates contractional deformation associated with closure of the oceanic tract between the island arc(s) of the Notre Dame Subzone and Laurentia (van Staal, 1987; van der Pluijm *et al.*, 1990, 1993; van der Voo *et al.*, 1991). This oceanic tract closed in a southeast-dipping (present coordinates) subduction zone, the remnants of which are located on the present northwestern side of the arc(s) of the Notre Dame Subzone (Fig. 1). Southeastward directed thrusting in the northern Exploits Subzone is therefore related to post-Taconic contractional events in the remaining open part of the Iapetus ocean, but, as emphasized by C. van Staal (pers. comm., 1995), simple temporal sequences need not be spatially related. The importance of this point is underscored by the late Arenigian (circa 480 Ma) thrusting of Exploits Subzone strata over Gander Zone rocks in the southern Exploits zone (Coleman-Saad *et al.*, 1992). This relationship ties the Exploits subzone to Gondwana during the Lower–Middle Ordovician, at the same time that the Notre Dame Subzone arcs were being accreted to the Appalachian margin of Laurentia (Fig. 18). However, paleogeographic reconstructions of where the Gander Zone rocks were deposited along the Gondwanan margins are poorly constrained, allowing deposition anywhere between the Avalonian margin and the Peruvian promontory (this uncertainty is delineated by the large circles along the Gondwana margin in Fig. 18).

SILURIAN–DEVONIAN STRUCTURAL EVOLUTION

The Fog Fault is offset along a set of dextral, north-northeast striking brittle faults (Fig. 4) that are demonstrably pre-Late Silurian, as they are locally intruded by dikes from the Hodges Hill Batholith. In other places, these faults cut the Hodges Hill Batholith (Fig. 2), suggesting that the batholith intruded during Late Silurian dextral strike-slip faulting. Locally, good exposure of these faults coupled with offsets of distinctive marker horizons along the north end of New Bay Pond demonstrates their dextral offsets, with separations ranging from a few centimeters to approximately 1/2 km on individual faults (Fig. 8). This sense of offset is also supported by stepping directions on numerous fibrous slickensides that trend parallel to the faults and plunge gently southwest and northeast (Fig. 16d), consistent with a slightly oblique strike-slip fault system, including uplift of the blocks to the west. A total dextral displacement of 3 km is mapped for these faults across the width of the map area (Fig. 4), although minor sinistral offsets (totalling 0.5 m) were also noted. Lafrance (1989) has similarly described the Boones Point Complex along the contact between the Roberts' Arm belt and Exploits Subzone (Fig. 2) as a Middle–Late Silurian dextral transpressional zone formed during the progressive steepening of beds (Fig. 17). Fabrics within the Boones Point Complex overprint older cleavages and folds (Lafrance, 1989). Undeformed Late Silurian granodiorite intrudes the Boones Point Complex (G.R. Dunning, pers. comm., 1985, cited in Lafrance, 1989), suggesting that dextral motion

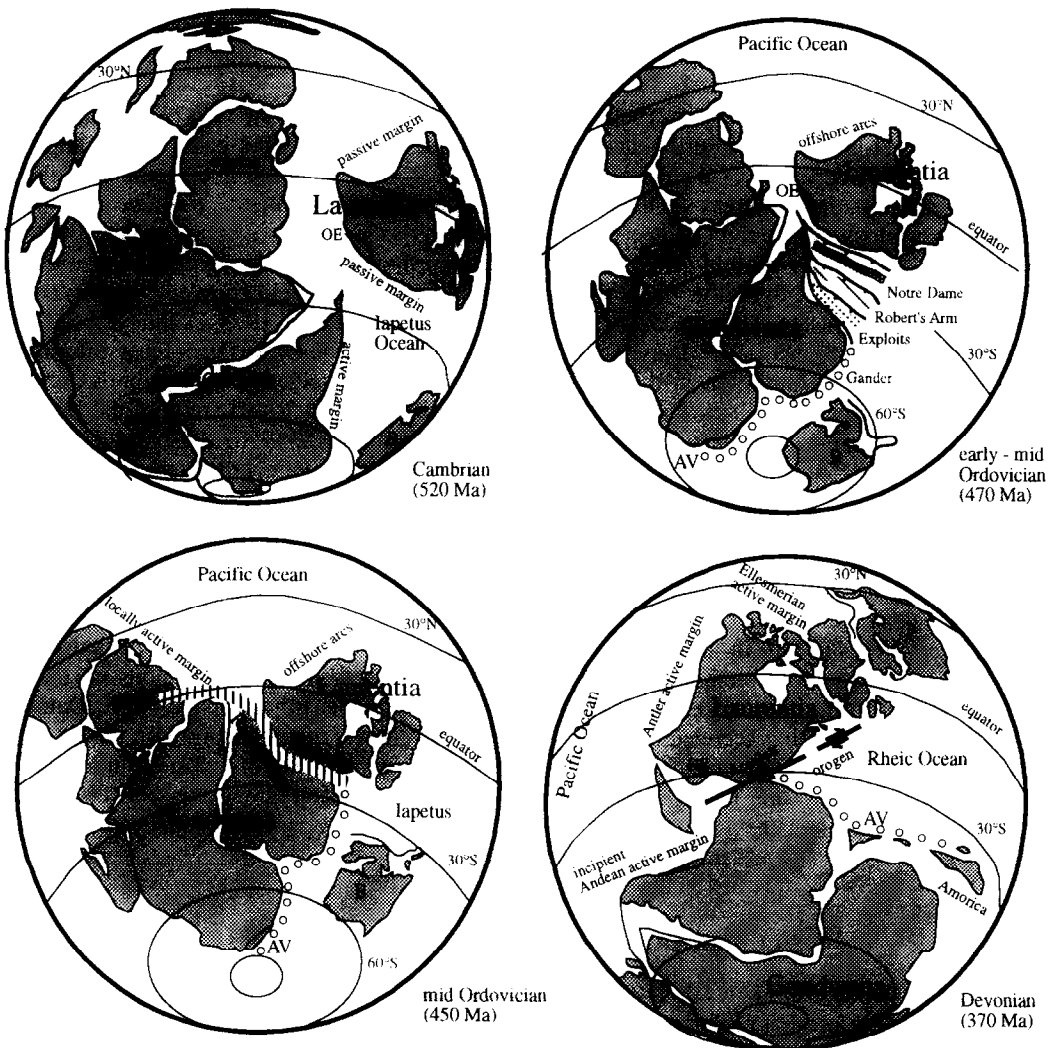


Fig. 18. Paleozoic paleogeographic reconstructions, based largely on Dalziel *et al.* (1994). The Cambrian Iapetus Ocean began closing by Early–Middle Ordovician, forming at least three arcs including the Notre Dame Subzone, adjacent to North America, the Roberts Arm belt, near the middle of the 4000 km wide ocean, and the Exploits Subzone, near the Gondwanan margin. The Exploits Subzone began being thrust over the Gander Zone in the Early Ordovician (Arenigian) during the Penobscottian orogeny, with deformation culminating in the Middle Ordovician (Carodocian). Continental margin rocks of the Gander Zone are represented by large open circles, and their wide distribution reflects the uncertainty about what part of Gondwana the Gander Zone was originally deposited. The Iapetus Ocean between Laurentia and Gondwana had largely closed by Middle–Late Ordovician, and Silurian–Devonian tectonism was dominated by the counterclockwise rotation of Gondwana, dextral transpression, and convergence between the Avalonian margin of Gondwana and the collisionally modified margin of Laurentia. Subduction zones likely existed beneath both the Avalonian margin of Gondwana and the Northern Appalachian margin of Laurentia at this time (Bradley, 1983). Paleogeographic maps are based largely on Dalziel *et al.* (1994), modified with data from van der Voo (1988) and van der Voo *et al.* (1991), and interpretations from this study.

ended by Late Silurian along this particular zone. Elliott *et al.* (1991) recognize that dextral ductile faulting on New World Island preceded intrusion of the 408 ± 2 Ma Loon Bay igneous suite, but also show that the Loon Bay suite is locally offset sinistrally on post-Silurian faults. Williams and Urai (1989) found dextral motion along the Indian Islands fault zone, which merges with and forms a small angle with the Reach Fault system. Kusky *et al.*, (1987) suggested dextral motion on the entire Northern Arm–Reach Fault system in Silurian and Devonian times.

Several of the Silurian subaerial/shallow marine overlap sequences in central Newfoundland (the circa 430–420 Ma Botwood and Springdale Groups: Williams, 1967; Chandler *et al.*, 1987; Whalen *et al.*, 1987; Elliott *et al.*, 1991) have been interpreted as deposits of pull-apart basins related to this Silurian dextral strike-slip fault system (Kusky *et al.*, 1987; Buchan and Hodych, 1992), and alternatively as subaerial thrust-related deposits (Karlstrom, 1982; Karlstrom *et al.*, 1982, 1983b). Problems with the thrust-related interpretation include the observation that these sequences are almost invariably bounded by strike-slip faults, not thrust faults, and such an abundance of volcanic rocks as is found in these basins is atypical of thrust-related foredeeps. A problem with the pull-apart interpretation is that some of the bounding strike-slip faults, such as the Northern Arm Fault, cut plutons that intrude rocks of the inferred pull-apart basins and therefore, the present day strike-slip faults are younger than the inferred deposits of the pull-apart basins. We interpret these relationships to indicate that motion on the bounding strike-slip faults was long-lived, and continued throughout and after extension in and filling of the pull-apart basins. It is typical to find extensional basins cut by contractional or translational structures along major strike-slip fault systems (Wilcox *et al.*, 1973; Mann *et al.*, 1983). The observation that the faults locally cut the plutons, and the plutons locally cut the faults, indicates that dextral slip continued throughout the closure of the Iapetus Ocean. Some of the Siluro-Devonian plutons, such as the Mount Peyton Pluton, and the Loon Bay Batholith, are located in dilational bends in these strike-slip fault systems, suggesting that Silurian fault systems also localized the site of pluton emplacement (Hutton, 1982; Hutton and Reavy, 1992).

CONCLUSIONS AND IMPLICATIONS

In summary, the southern Exploits Subzone began being thrust over Gander Zone continental margin rocks in late Arenigian. Caradocian–Early Silurian thrusting and imbrication of the northern Exploits Subzone formed a south-vergent accretionary complex, in response to the collision of the Exploits Subzone arc with the Gondwanan continent (Fig. 18). The linkages of the Exploits Subzone with Gondwana, and the Notre Dame Zone with Laurentia, by Early–Middle Ordovician, plus the paleomagnetically-documented paleolatitudinal separation of the Notre Dame and Roberts Arm belts in the Middle Ordovician, suggests that closure of Iapetus in Middle Ordovician times involved at least three subduction zones (Fig. 18). These included (1) a south-dipping zone beneath the Notre Dame Zone, along which the Taconic allochthons were emplaced; (2) a south-dipping zone beneath the Roberts Arm belt, which by Middle Ordovician times led to the collision of the Roberts Arm belt and the Notre Dame Zone; and (3) a south-dipping zone beneath the Exploits subzone. A marginal basin separated the Exploits Subzone from Gander Zone sediments deposited on the Gondwanan continental margin, and this marginal basin closed in Middle Ordovician times, generating the post-Caradocian clastic wedge and south-vergent fold-thrust structures in the Exploits Subzone.

Silurian–Devonian tectonism was dominated by dextral transpression (Dewey, 1982; Kusky *et al.*, 1987; Dalziel *et al.*, 1994), although rifting events within the broader context of the

counter clockwise rotation of Gondwana relative to Laurentia may have separated fragments and then later brought them together at different locations along the transpressional orogen (Dalziel *et al.*, 1994). For instance, the Devonian coastal volcanic arc built on Avalonian crust (Bradley, 1983) above a south dipping subduction zone collided with a contemporaneous arc now preserved in the Bronson Hill–Boundary Mountains anticlinorium (Piscataquis arc), formed above a north-dipping subduction zone descending beneath the amalgamated Notre Dame Subzone, Roberts Arm belt, and Exploits Subzone (Bradley, 1983). These different terranes may have been left attached to North America by an outboard jumping of the locus of dextral strike-slip faulting, or by a Siluro-Devonian rifting event. In either case, great uncertainty remains about on which of the margins of Gondwana sediments of the Gander Zone were deposited. With present data, these sediments could have been deposited anywhere between the Avalonian terranes adjacent to northwest Africa, and the Peruvian Promontory (Fig. 18).

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