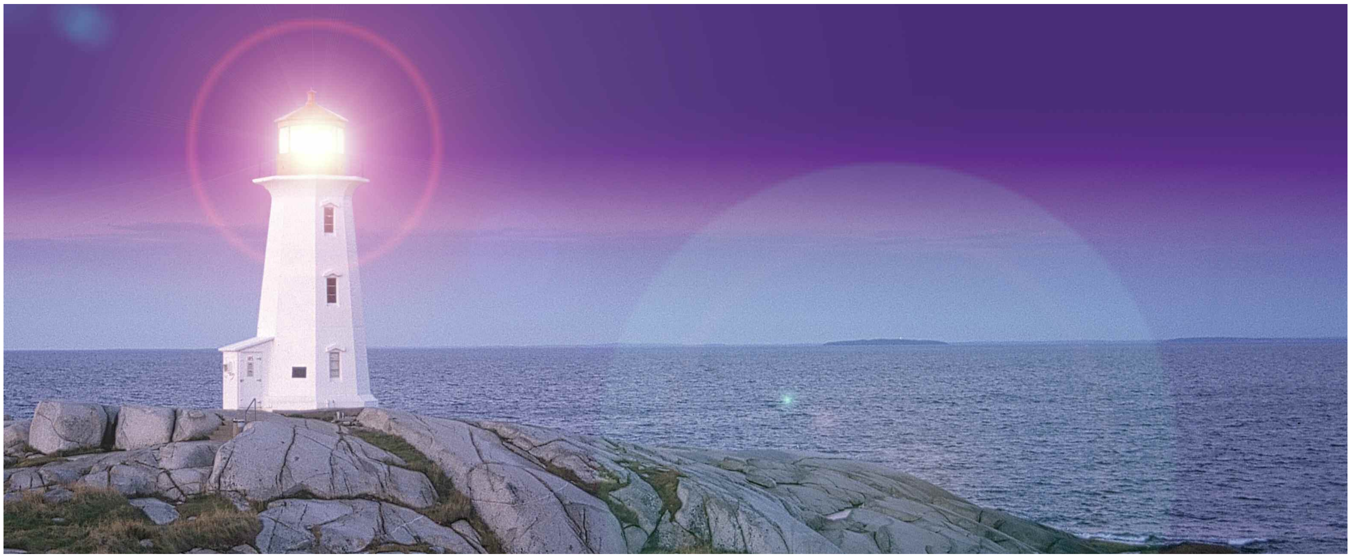
GAC-MAC-CSPG-CSSS

HALIFAX 2005

Building Bridges—across science, through time, around the world



Field Trip B7

Transpression and transtension along a continental transform fault: Minas Fault Zone, Nova Scotia

John W.F. Waldron, Joseph Clancy White, Elizabeth MacInnes, and Carlos G. Roselli

Geological Association of Canada
Mineralogical Association of Canada - Canadian Society of Petroleum Geologists - Canadian Society of Soil Sciences
Joint Meeting - Halifax, May 2005

Field Trip B7

Transpression and transtension along a continental transform fault: Minas Fault Zone, Nova Scotia

John W.F. Waldron[1], Joseph Clancy White[2],
Elizabeth MacInnes[2], and Carlos G. Roselli[1]

[1] *Department of Earth and Atmospheric Science, University of Alberta, Edmonton, Alberta, Canada T6G 2E3*
[2]*Department of Geology, University of New Brunswick Fredericton, New Brunswick, Canada E3B 5A3*

Department of Earth Sciences
Dalhousie University
Halifax, Nova Scotia, Canada B3H 3J5

ISBN 0-9737982--2-X
AGS Special Publication Number 33

SAFETY INFORMATION

The coastal exposures of Nova Scotian geology offer spectacular views but also some very real hazards. For personal and group safety we ask all participants to read and heed the following safety-related procedures. We ask for your cooperation and common sense in making this a safe and enjoyable field trip for everyone. Thank you.

CLIFFS AND FALLING ROCKS: Many of the cliffs on the Bay of Fundy shore undergo active erosion; cliff falls may occur at any time. Avoid approaching vertical or overhanging cliffs, and wear a HARD HAT when working close to any cliffs. Do not climb the cliffs or piles of talus. If you have to move high on any slope, watch for people below you.

TIDES: The Minas Basin enjoys the largest tidal range in the world (up to 16 m). On areas of low gradient - mud and sand flats - the water level rises with a speed such that **it is not possible to outrun the rising tide.** In addition, there are coves where it is possible to be trapped by the rising tide with no escape route. Consult tide tables or tide times published daily in most provincial newspapers before visiting them. For most coastal sections, it is best to work during the falling tide. If you work past low tide, keep an eye on your escape route. Do not venture onto sand flats or mud flats.

INTERTIDAL AREAS: Avoid muddy areas, especially in the lower intertidal zone, where you can become stuck during a rising tide. Intertidal rocks can be exceedingly slippery, resulting in a swift and unforgiving fall. Avoid damp rocks and in particular those with green algal growth (*very* slippery).

SUITABLE CLOTHING: Participants should have adequate footwear and protection against both wet and cold, including a hat, gloves, and boots. Adequate clothing is important if you are involved in an accident or if you are required to spend an extensive period of time outdoors. Spring weather in Nova Scotia is unpredictable and can change from sunny and warm, to rain, wet snow, and high winds with little notice.

MINE WORKINGS: When visiting quarries and mines, it is paramount that all required safety equipment is properly worn and that all rules be strictly followed. Stay with the group during the tour.

ROCK HAMMERS: Use caution when hammering: be aware of people around you, use controlled downward blows, and do not hammer indiscriminately. When hammering, either shield your eyes or wear protective eyewear. Do not hammer near other members of the group. Use rock hammers only; a carpenter's hammer may splinter and send metal chips flying. If using a chisel, please ensure it is approved to be used as such. Never use a second rock hammer as a chisel. Gloves are recommended when using a rock hammer.

ROADS: Most of the roads in the area are well travelled and traffic may travel at high speed. Do not venture onto the road unless you are crossing the road, and only cross with the group to minimize traffic disruption. We will try to park safely off the road on the same side as the outcrops you are visiting. If this is not possible, pay careful attention to your own safety when crossing. Bear in mind that drivers are not used to seeing pedestrians in most of the locations we visit.

VEHICLE TRAVEL: Remain seated when the vehicle is in motion. All knapsacks, rock hammers, rock samples etc. should be safely stowed underneath your seat. Nova Scotia law requires that you wear a seat belt at all times.

FIRST AID / MEDICAL CONDITIONS: Several First Aid kits will be located on the vehicles. Co-leaders of the field trip will carry first-aid kits Field trip participants with medical conditions (allergies, diabetes, etc.) may wish to advise the field trip leaders prior to departure. All personal medical information will be treated with strict confidence.

IN THE UNLIKELY EVENT OF AN EMERGENCY, CALL *911*. Some of the region has poor cellular phone coverage, so it may be necessary to use a pay/private phone.

ACKNOWLEDGMENTS

We are grateful to numerous students and field assistants who have carried out fieldwork for us during our work on the Minas Fault System. This work was made possible by NSERC Discovery Grants to Waldron and White. Sandra Barr, and the organizing committee of the 2005 GACMAC annual meeting provided support and encouragement during preparation of the guide. Brendan Murphy provided helpful editorial comment. Sydney Lancaster assisted with the assembly and editing of the manuscript.

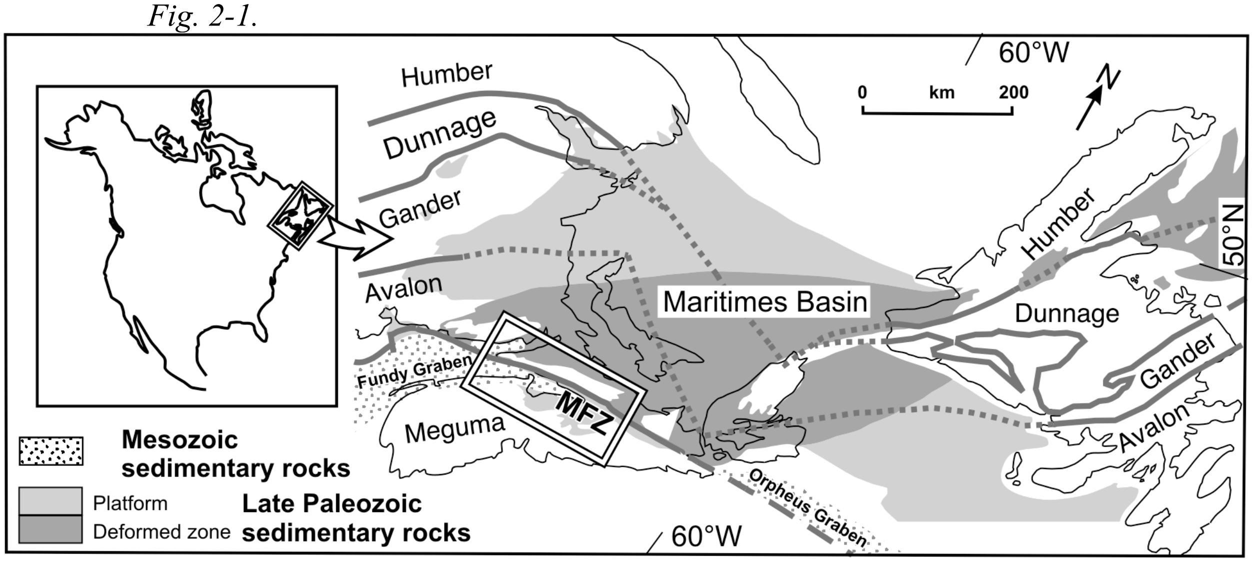
TABLE OF CONTENTS

1. INTRODUCTION

In the Northern Appalachians of Atlantic Canada, a major fault zone, the Minas Fault Zone (MFZ), strikes east-west through Atlantic Canada (Fig. 1-1). The zone was tectonically active at intervals throughout the late Paleozoic and early Mesozoic eras. The boundary is marked by a conspicuous series of topographic lineaments, and is visible from space. Its lateral extent and linear trace, which approximates a segment of a great-circle, suggest that it was a transform fault, active during late stages in the development of the Appalachians. The MFZ separates two terranes of the Appalachian Orogen, the Meguma Terrane to the south and the Avalon Terrane to the north. These terranes have strikingly different Early Paleozoic histories but show related stratigraphies from the Devonian onward. Nonetheless, the MFZ continued to exert a strong control over the history of sedimentation, basin development, and deformation during the Carboniferous and into the Mesozoic. The focus of this field trip is the history of deformation along the MFZ, as displayed in spectacular coastal outcrops on the shores of the Minas Basin, the innermost portion of the Bay of Fundy.

On the north shore of the basin, transposed mylonitic rocks, breccias, and mélanges dominate an east-west belt of outcrop along the MFZ. Immediately south of the boundary, on the shores of the Minas Basin of the Bay of Fundy (Fig. 2-1), exceptional exposures display systems of folds, including overturned and downward-facing structures; faults are also abundant, and include examples with extensional, contractional, and strike-slip offsets. Seismic and drilling data from petroleum exploration imply duplication of stratigraphy above a zone of Mississippian evaporites. This guide describes and interprets these structures as classic examples of transpressional deformation, and shows how they provide important constraints on the later stages of tectonic development of the Appalachian orogen.

Fig. 1-1: *Major elements of the geology of the Canadian Appalachians. Box encloses area of Fig. 2-1.*



2. GEOLOGIC SETTING

ELEMENTS OF THE APPALACHIAN OROGEN IN ATLANTIC CANADA

Appalachian Orogen

Nova Scotia lies within the Appalachians, an orogen that stretches from Alabama to Newfoundland. Extensions of the Appalachian belt occur in the Ouachitas of southeastern USA, and in the Caledonides of the British Isles and Scandinavia. The units that comprise the Appalachian orogen were assembled as a result of the progressive convergence and transcurrent motion between Laurentia and Gondwana-derived terranes during the Paleozoic, and subsequently affected by extension during the break-up of Pangea in the early Mesozoic.

Numerous episodes of orogenic activity have been identified in the Appalachians. These include the development of the Laurentian margin and ophiolite obduction during the Mid-Ordovician Taconic orogeny (~490-470 Ma), the accretion of peri-Gondwanan terranes during the Silurian Salinic orogeny (~ 435-420 Ma), the deformation of the Meguma and Avalon terranes during the Devonian Acadian orogeny (~410-380 Ma), and the formation of Pangea by Laurentia-Gondwana collision during the Permo-Carboniferous Alleghenian-Variscan orogeny (~280 Ma).

Northern Appalachian zones and terranes

Several schemes have been proposed in which the Appalachians are divided into zones and/or terranes, interpreted as representing separate pre-Appalachian continental margins, continental fragments, and island arcs, etc (e.g. Williams 1979, 1995, Williams et al. 1988). The major elements of one simple scheme are shown in Fig. 1-1. The zones recognized are:

- Humber Zone: This zone represents the Cambrian to earliest Ordovician passive margin of Laurentia, overlain by foreland basins developed during Middle Ordovician to Devonian time, though much of the foreland basin record is hidden under the Gulf of St. Lawrence.

- Dunnage Zone: The Dunnage Zone represents fragments of the Iapetus Ocean. Williams et. al (1988) recognized that the Dunnage zone is composite: northwestern parts of the zone represent oceanic crust that was emplaced above the Laurentian margin in Early Ordovician time. In contrast, southeastern parts of the zone record early interaction with continental fragments of Gondwana.

- Gander Zone or Terrane. The Gander Terrane is characterized by metamorphic rocks in facies from greenschist to (locally) blueschist facies representing continental fragments of Gondwana to the southeast of the main Iapetus Ocean. Numerous subdivisions have been proposed.

- Avalon Zone or Terrane. The Avalon Terrane is characterized by low-grade metamorphic, igneous, and sedimentary rocks. Abundant late Neoproterozoic igneous rocks record tectonic activity on a margin of Gondwana, from which the Avalon Terrane was rifted, probably in Cambrian time. Paleozoic cover rocks record shelf sedimentation in Cambrian and Silurian time in various parts of the terrane.

- Meguma Zone or Terrane. The Meguma Terrane is the most outboard of the Appalachian terranes, and is exposed in southern mainland Nova Scotia. The Meguma Terrane is principally characterized by a thick (>10 km) succession of Cambrian to Ordovician turbiditic meta-sandstones and slates (the Meguma Group) overlain by thinner Silurian to Devonian volcanics and shelf sediments.

Maritimes Basin

These elements were assembled into something like their present-day arrangement by the Middle Devonian Acadian Orogeny. From this time until the late Triassic or Early Jurassic, Atlantic Canada lay near the centre of Pangea. However, the area remained tectonically active. In the Late Devonian to at least the Early Permian a very large sedimentary basin, the Maritimes Basin (Fig. 1-1), developed. The Maritimes Basin is a composite basin, superimposed across the collage of assembled Appalachian terranes, extending from the Humber Zone in the north to the Meguma in the south. Thick deposits of clastic, carbonate, and evaporites filled this basin. Portions of the basin show clear evidence for tectonism that continued through deposition, including both transpression and transtension. This Late Paleozoic deformation will be the focus of the trip.

Mesozoic rifting

In Triassic time, the earliest effects of the break-up of Pangea were manifested between eastern North America and northwest Africa. An extensive rift system was filled with clastic sediments, local evaporites, and mafic volcanics, collectively assigned to the Newark Supergroup. A large half-graben, the Fundy Basin, occupies the site of the present Bay of Fundy, and its fill (Fundy Group) extends onto both the north and south shores of the Minas Basin (Fig. 1-1).

AVALON TERRANE IN NOVA SCOTIA

The Avalon Terrane originated as one of several arc-related terranes that formed along the periphery of western Gondwana in the Neoproterozoic (O'Brien et al. 1983; Johnson and Van der Voo 1986, Nance and Murphy 1994, Keppie et al. 1996, O'Brien et al. 1996, Murphy et al. 1997). The Lower Cambrian is characterized by volcanics and continental to shallow marine sediments also related to dextral strike-slip faulting (Murphy et al. 1985, 1991). By late Ordovician or Early Silurian, Avalonia was apparently in collision with Laurentia (e.g. Murphy et al. 1996). A fossiliferous Silurian succession records rapid latest Silurian subsidence (Waldron et al. 1996) interpreted to represent loading resulting from overthrusting of the Meguma Terrane to the south marking the onset of the Acadian orogeny. Acadian deformation in the Avalon Terrane produced complex overprinting relationships. Uplift and erosion occurred between 380 and 365 Ma (Murphy et al. 2000, Murphy and Hamilton 2000).

Fountain Lake Group volcanic rocks were extruded in the Devonian, immediately following the main Acadian orogeny. They have an intraplate signature and are associated with crustal extension (Piper et al. 1999*)*. The Fountain Lake Group was tilted to near vertical prior to intrusion of the earliest Carboniferous plutons (Piper 1994). During the Middle Tournaisian to Middle Visean, north- to northwest-vergent thrusting was synchronous with pluton emplacement and half-graben formation (Piper and Pe-Piper 2001). Transpression continued until at least mid-Carboniferous, as recorded by thrusts in southeastern New Brunswick (Plint and Poll, 1984) and by mylonites in the Cape Chignecto pluton (Waldron et al. 1988). To the north and east, the Cobequid Highlands are unconformably overlain by Pennsylvanian Cumberland Group clastic sediments. However, along the south edge of the Cobequid Highlands, the MFZ includes fault-bounded slivers of Cumberland Group and Mesozoic rocks, indicating that deformation continued into at least the Jurassic.

MEGUMA TERRANE IN NOVA SCOTIA

The Meguma Terrane is characterized by a thick (>10 km?) succession of Cambrian-Ordovician turbidites, the Meguma Group. These have been interpreted as submarine fan deposits (Schenk 1970, 1991; Waldron & Jensen 1985, Waldron 1992) deposited on a margin of Gondwana. They are overlain by a Middle Ordovician to Silurian succession of bimodal within-plate volcanics and siliciclastics (Whiterock Formation), and in turn overlain by Early Devonian limestones and clastics (Torbrook Formation).

The nature of the unexposed basement to the Meguma Group is speculative; records of the original relationship between the Avalon and Meguma terranes has been destroyed by displacement along the MFZ. Poorly exposed high-grade metamorphic rocks of the Liscomb Complex, exposed in dome-like structures within the Meguma Terrane, may include material derived from basement, but it is uncertain whether the basal contact of the Meguma Group is a depositional contact or a thrust.

Regional deformation during the Acadian orogeny is recorded by folding, cleavage development, and low-grade metamorphism from 415-377 Ma (Keppie & Krogh 1999). Northeast-trending upright horizontal folds have been proposed to be associated with the docking of Meguma with Laurentia (Henderson et al. 1986, Muecke et al. 1988, Culshaw & Liesa 1997) while axial stretching during folding suggests concomitant orogen-scale transpression (Culshaw & Liesa 1997).

Deformation was followed by voluminous granitoid plutonism (380-370 Ma). Dextral shear zones (ca. 375 Ma) display S-C fabrics, rotated porphyroblasts and sheath folds (Keppie et al. 1991, Murphy & Keppie 1998). Rapid uplift and erosion occurred between 370 and 360 Ma to expose plutonic and metamorphic rocks (Keppie & Dallmeyer 1995). The earliest rocks of the Maritimes Basin rest unconformably on all these older units, and are assigned to the Horton Group (?Upper Devonian to Tournaisian). However, in some areas, structures in the Meguma Group were reactivated and amplified by deformation that continued well into the Carboniferous (Culshaw & Liesa 1997).

HISTORY OF THE AVALON-MEGUMA BOUNDARY

The Minas Fault Zone is over 300 km in length and defines the boundary zone between the Avalon and Meguma terranes. This terrane boundary has been referred to as the Cobequid-Chedabucto fault zone (CCFZ) (Webb 1969), the Minas Geofracture (Keppie 1982), the Cobequid-Chedabucto fault system (Mawer and White 1987), and the Minas fault system (Gibbons et al. 1996).

The Early Paleozoic relationships between Meguma and Avalon, across this zone, are controversial. Although both terranes have general affinities with Gondwana, their contrasting Cambrian-Ordovician history suggests that their tectonic environments were strikingly different. Schenk (1970) argued that the Meguma Terrane originally bordered NW Africa and was transferred to Laurentia during the Acadian orogeny. Paleocurrent, geochronologic, and isotopic evidence supports the link with Gondwana at least in the Cambrian (Krogh and Keppie 1990, Waldron 1992). There are indications from seismic and isotopic studies that the Meguma Terrane have been thrust over Avalon (Eberz et al. 1991, Keen et al. 1991).

However, more recently other authors have suggested that the Meguma Terrane that Avalonia and Meguma have been contiguous back to the late Neoproterozoic and represent a single fragment of Gondwana (Keppie & Krogh 1999, Murphy et al. 2004). For example, Upper Ordovician to Lower Silurian clastic rocks in the Meguma Terrane contain detrital zircon populations consistent with an Avalonian source, while the chemistry of contemporary within-plate volcanic rocks shows evidence that the underlying continental basement had Avalonian ε_{Nd} characteristics (Keppie & Dostal 1980, Keppie et al. 1997)

Regardless of their precise provenance, early Paleozoic strata of the Meguma Terrane were extensively shortened during the Acadian Orogeny, which produced a regional system of folds, slaty cleavage, and metamorphism (Henderson et al. 1986). During later stages of Acadian deformation, structures display an increasing prominence of dextral strike-slip. For example, S-C protomylonites and other structural features in granitoid rocks close to the MFZ indicate predominantly dextral strike-slip motion during the later stages of Acadian deformation (Eisbacher 1969, 1970, Mawer & White 1987). Deformation intensity within the post-Acadian overstep sequences along the Avalon-Meguma boundary is highly asymmetric. The northern boundary with Avalonia is characterized by steeply dipping, high-strain-gradient deformation zones, while the southern boundary with Meguma exhibits a distinctively more distributed deformation. Despite the latter contrasts, the kinematic history recorded by structures is strikingly similar.

MARITIMES BASIN

The Maritimes Basin extends across all five zones of the Appalachians, and contains, at its centre, up to twelve kilometres of Late Devonian to Early Permian sedimentary fill (e.g. Durling & Marillier 1993). Late Devonian and Early Carboniferous basin development has been interpreted as related to major strike-slip motion within the Appalachians (Bradley 1982), including clockwise rotation of Laurentia relative to Gondwana (Keppie 1982, Keppie et al. 1996, Murphy 2000). The crust beneath the basin is significantly thinned (Marillier & Verhoef 1987), and the basin has many rift-like characteristics (McCutcheon & Robinson 1987, Gibling

1995). The fill is not evenly distributed; the basin contains numerous intracontinental depocentres and sub-basins that developed during the final stages of the Acadian orogeny (Keppie 1982, St. Peter 1993, Murphy et al. 1995, 1999a, Gibling 1995).

A broad central zone, elongated NE-SW and extending from southern New Brunswick to southwestern Newfoundland (Fig. 1-1), contains successions that are thicker, and generally more deformed, than those outside this zone (Gibling 1995). The areas outside this central zone are sometimes characterized as 'platformal' although they nonetheless contain significant facies and thickness changes. The areas to be visited by this field trip lie just outside the central thick zone, but include perhaps the most intensely deformed rocks in the entire Maritimes Basin.

The fill of the basin is complex, and stratigraphic subdivision has often been confusing. Significant rationalization of the stratigraphy has been achieved since 1990 (e.g. Ryan et al. 1991). In most areas of the basin, a simple subdivision into six groups has been adopted (Table 2-1), and this subdivision will be followed in this guide.

Table 2-1: *Major stratigraphic units in the Maritimes Basin*

Group	Typical lithologies	Age range
Pictou Group	Predominantly red, non-marine clastic sedimentary rocks	Westphalian D to Stephanian (to early Permian?)
Cumberland Group	Predominantly grey, coal-bearing, non-marine clastic sedimentary rocks	Late Namurian to Westphalian D
Mabou Group	Predominantly red, generally fine-grained lacustrine clastic sedimentary rocks, minor non-marine evaporites	Late Visean to Middle Namurian
Windsor Group	Marine to restricted limestones, evaporites, and predominantly fine-grained red clastic sedimentary rocks	Visean
Horton Group	Grey to red, mainly lacustrine to fluvial clastic sedimentary rocks	Late Devonian to Tournaisian
Fountain Lake Group	Bimodal volcanics and volcaniclastic rocks and associated, mainly red, subaerial clastic sedimentary rocks.	Late Devonian

The episodic nature of dextral strike-slip displacement during the Upper Devonian and Carboniferous resulted in a complex interplay between compressional and extensional tectonics, resulting in sedimentation in one basin and synchronous deformation in another (Keppie 1993, Murphy et al. 1995, Gibbons et al. 1996, Murphy & Keppie 1998). South of the Cobequid fault segment of the MFZ, the Horton Group was deposited in active SW-NE striking half-grabens (Martel & Gibling 1996, Murphy et al. 1995).

This initial fault-controlled rapid subsidence was followed by slower, more generalized subsidence during the later Carboniferous (Bradley 1982, McCutcheon & Robinson 1987). However, subsidence of the basin was interrupted by intense deformation along the MFZ in mid-Carboniferous time. In contrast to other areas of the basin, a profound angular unconformity separates intensely deformed Horton, Windsor, and Mabou Group rocks from the overlying, more gently deformed Cumberland and Pictou Groups.

Complex patterns of deformation persisted during the Pennsylvanian (Late Carboniferous). North of the Minas Fault zone, Windsor Group evaporites were mobilized during this interval; the rise of evaporite walls, and the simultaneous subsidence of minibasins due to evaporite withdrawal, controlled sedimentation in the Cumberland sub-basin (Waldron & Rygel 2005). Farther east, dextral strike-slip motion opened a pull-apart structure, the Stellarton sub-basin, at a releasing bend in the Cobequid-Hollow fault system, causing rapid subsidence of a deep basin. Deformation was synchronous with sedimentation but continued afterwards as the basin was transferred to a transpressional setting on another part of the fault system (Waldron 2004).

MESOZOIC HISTORY

A series of Mesozoic strike-slip, oblique-slip and normal faults bound the present-day Minas Basin. Many of these faults are interpreted as reactivated Paleozoic contractional structures (Olsen & Schlische, 1990). The Cobequid fault became an oblique-slip fault with normal and sinistral strike-slip components of displacement (Withjack et al. 1995).

The asymmetric Fundy half-graben contains several kilometres of non-marine sedimentary rocks and basalt flows assigned to the Fundy Group of the Newark Supergroup (Olsen & Schlische 1990, Withjack et al. 1995). These Mesozoic synrift units include the Wolfville and Blomidon Formation (Middle to Late Triassic clastic sedimentary rocks); the Early Jurassic tholeiitic North Mountain Basalt; and the overlying clastic sedimentary rocks of the Early Jurassic McCoy Brook Formation. Within the basin the strata thicken toward the north, indicating part of the MFZ was active during the Middle Triassic to Early Jurassic deposition of the synrift strata (Withjack et al. 1995).

Withjack et al. (1995) propose two distinct episodes of deformation during Mesozoic time. The first episode from Middle Triassic to Early Jurassic, was obliquely extensional. The rifting was associated with NW-SE extension which reactivated NE-trending Paleozoic compressional structures as normal faults and east-trending Paleozoic compressional structures as oblique-slip faults with normal and sinistral strike-slip components; these faults formed the northern boundary faults of the Fundy Basin. Basin inversion was associated with NW-SE shortening along the faulted margins of the Fundy Basin during or after Early Jurassic time and probably before or during Early Cretaceous time. The northwestern boundary faults experienced several kilometres of reverse displacement, broad anticlines developed within their hanging walls, and the Fundy Basin acquired its synclinal form. The shortening resulted from tectonic adjustments associated with seafloor spreading in the Atlantic (Withjack et al, 1995).

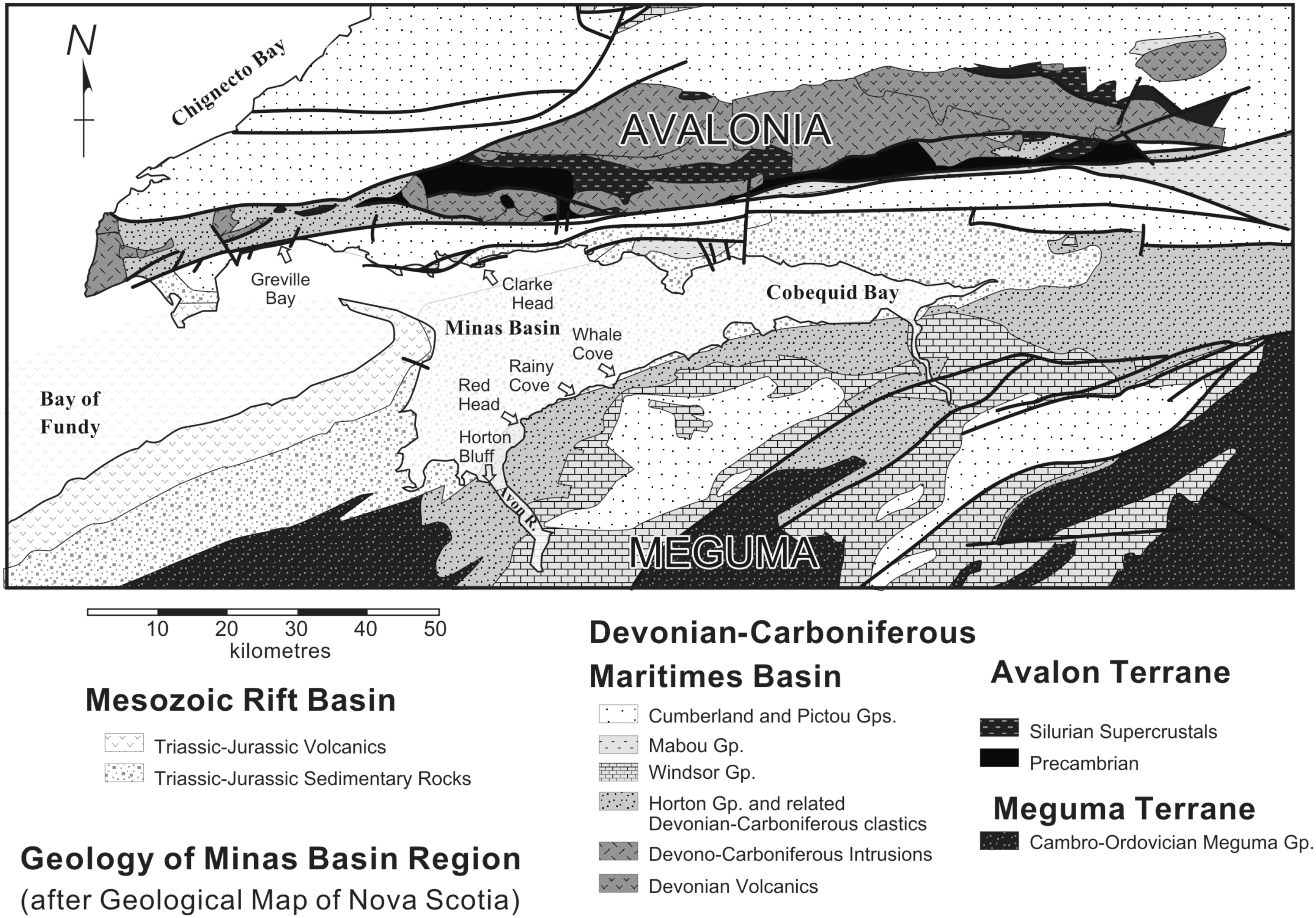


Fig. 2-1: *Geological map of the Minas Basin area, Nova Scotia, modified from Donohoe and Grantham (1989).*

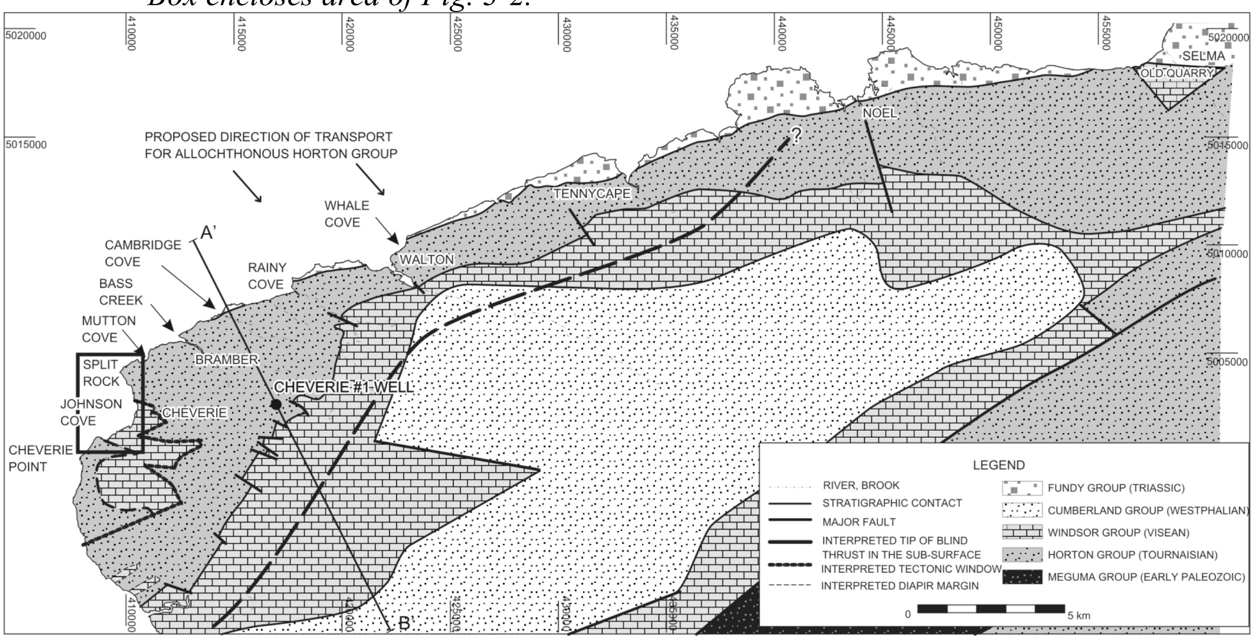
3. MINAS BASIN: SOUTH SHORE

OVERVIEW OF THE GEOLOGY

Previous mapping

Because of its accessibility, spectacular coastal exposure, and the presence of significant mineral resources, the south shore of the Minas Basin has been the subject of numerous mapping efforts, though the patchwork of maps with varied objectives and scales makes regional synthesis difficult. Various parts of the area of study were mapped by the Geological Survey of Canada (Weeks 1948, Bell 1960, Boyle 1957, Stevenson 1958, Crosby 1962) and the Nova Scotia Department of Natural Resources (Ferguson 1983, Giles & Boehner 1982, Moore 1986, 1993a, 1993b, 1994, 1996, Moore & Ferguson 1986, Moore et al. 2000). Fig. 3-1 represents a summary of the principal features of the geology.

***Fig. 3-1**: Simplified geological map and cross-section of the south shore of the Minas Basin. Box encloses area of Fig. 3-2.*



Field relationships around Cheverie

The Horton Group in this area is divided into a lower, mainly grey lacustrine unit named the *Horton Bluff Formation*, overlain by predominantly fluvial red and grey beds of the *Cheverie Formation*. Relatively undeformed strata characterize the type sections of the Horton Bluff and Cheverie Formations, respectively west and east of the Avon estuary (Fig 2-1). The overlying Windsor Group has complex internal stratigraphy, but we shall see only the lowest units. A thin, basal unit of laminated lime mudstones, the *Macumber Formation*, is overlain in many places by enigmatic breccia units informally termed *Pembroke breccia*. Overlying gypsum and anhydrite are assigned to the *White Quarry Formation*.

At Cheverie Point (Fig. 3-2), the gently folded Cheverie Formation and overlying Macumber Formation plunge beneath a broad region of Pembroke Breccia. Scattered outcrops of highly deformed Windsor gypsum and anhydrite occur in adjacent coastal cliffs. The Cheverie Formation reappears farther northeast in antiform at Johnson Cove (Fig. 3-1, 3-2). On the north flank of this structure, the Cheverie Formation again dips north beneath Macumber limestone and Pembroke Breccia. However, immediately north, where a Windsor succession would be expected, is a region of coast within which Windsor and Horton lithologies are intermixed and have highly variable bedding orientations, interpreted as a megabreccia. Moore (1996) interpreted a north-dipping thrust fault through this zone. The thrust fault is overlain by Horton Bluff Formation, representing strata originally deeper in the Horton Group, exposed in a succession of ENE and WSW-trending folds with axial plane cleavage increasing in intensity north, as far as Split Rock (Fig. 3-2).

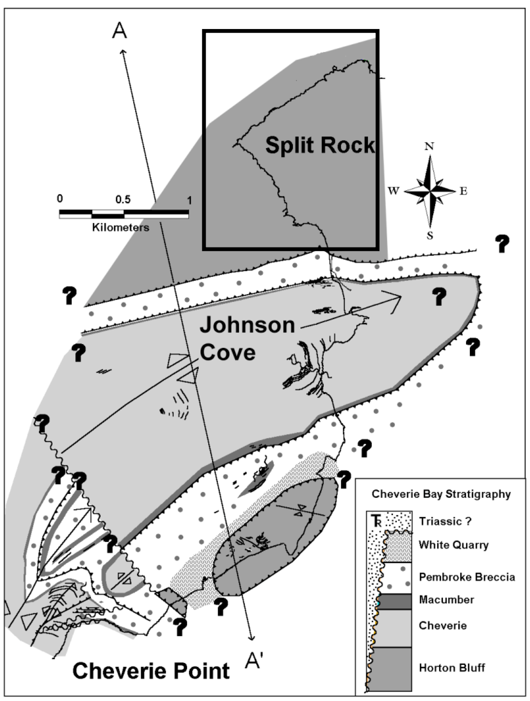
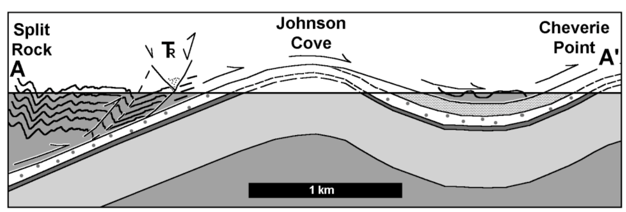


Fig. 3-2: *Schematic geologic map and cross-section of Johnson Cove and Split Rock, after Johnston (1999). Toothed lines indicate low angle faults; wavy lines are high-angle faults. Unornamented lines show trace of bedding. Box encloses area of Fig. 3-5.*



Relationships between Bramber and Walton

East of Cheverie, Horton Bluff Formation strata continue to be exposed in coastal cliffs, between regions of unconformable Triassic cover. Typically, the formation is deformed by ENE-trending folds, and cut by steep faults that strike northwest, typically marked by stream valleys.

Inland, the Horton Bluff Formation appears to pass stratigraphically upward into poorly exposed, steeply-dipping, generally south-younging Cheverie Formation. A belt of lower Windsor Group rocks, including Macumber Formation, marks the SE edge of the belt of Horton rocks. It is marked by sink holes (formed over gypsum) and outcrops of steeply-dipping carbonates, indicating that this belt represents the steep limb of a somewhat sinuous monocline. Extensive mineralization of the Macumber Formation at Walton, and elsewhere on this zone, has led to a history of drilling and mining operations.

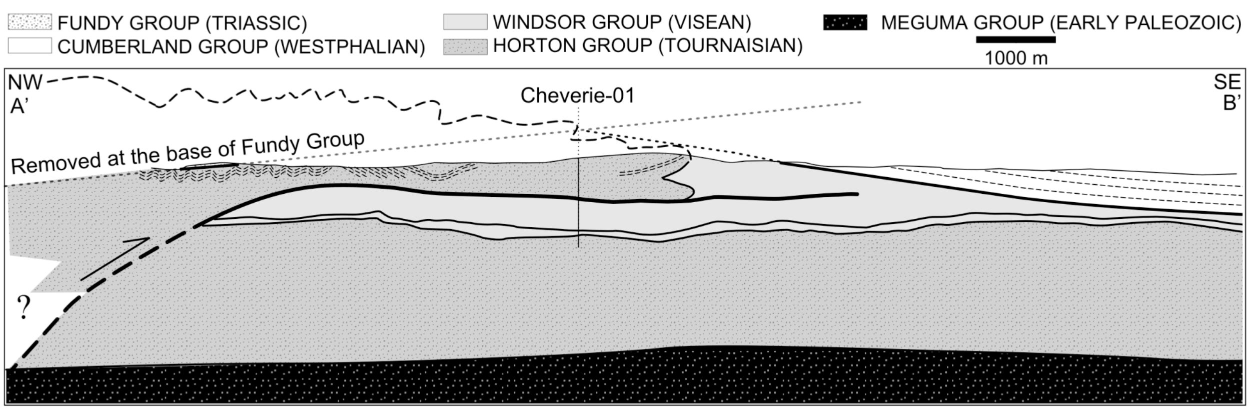
Regional structures

To the southeast of the coastal belt of highly deformed Horton Group rocks, the Windsor Group occupies a broad basinal area extending as far south as the Rawdon fault, which marks the north margin of a horst cored by Meguma basement. Around this basin, to the east and west, Horton and low Windsor Group rocks show relatively gentle inward dips. In contrast, stratigraphically higher units towards the middle of the basin are extensively deformed, with highly variable, steep dips. Outcrops in gypsum quarries (Ferguson 1983, Moore & Ferguson 1986) show folding and overturning. Although this central area of the Kennetcook basin is interpreted by Moore and Ferguson as a mosaic of steep block-faults, we infer from the decrease in deformation down-section that many of the faults actually sole into Windsor evaporites. At one point, the stratigraphic position of the evaporites is marked by a mega-breccia recorded by Moore and Ferguson (1986) lending support for the idea of a detachment at this level.

We interpret these relationships as indicating that the highly deformed coastal units of Horton Group were thrust over a more autochthonous succession of Horton Bluff, Cheverie, and Macumber Formations. The Pembroke Breccia, and the highly deformed evaporites on the coast of Cheverie Harbour, probably represent a thick deformation zone - effectively a décollement in Windsor evaporites - above which Horton Bluff Formation was transported.

To the southeast, in the central Kennetcook basin, there is no duplication of stratigraphy, though the structural contrast remains, so the thrust is interpreted to cut up-section southward in its hangingwall until it comes to lie between deformed Windsor Group (above), and relatively less deformed lower Windsor and Horton Groups (below). The hangingwall cut-off of the Cheverie Formation must lie between Johnson Cove and the central part of the Kennetcook basin; it is inferred to underlie the monoclinal structure inland just southeast of the Cheverie-01 well, that passes through the Walton mine (Locations shown in Fig. 3-3).

Fig. 3-3: *Schematic cross section along line A'B' (Fig. 3-1), after Roselli (2004).*



The inferred thrust does not outcrop within the area of Fig. 3-1. Folds in the hangingwall are likely detachment folds, which absorb some of the shortening, so it is likely that the thrust loses displacement to the south. Displacement may decline to zero within the area of Fig. 3-1, or farther south.

Previous authors have invoked thrusting in this area. Boehner (1991) interpreted Chevron Canada seismic lines as showing a low angle, north-dipping thrust fault, the Kennetcook thrust, with a mapped trace southeast of the Walton monocline. The interpretation favoured here is similar, but we infer that the Kennetcook thrust does not outcrop at the surface, and therefore is categorized as a 'blind' thrust.

HORTON BLUFF: HORTON BLUFF FORMATION TYPE SECTION

Access

From Halifax take highway 101 towards Wolfville. Take Exit 9 at Avonport.
0.1 km In Avonport turn immediately left.
0.5 km T junction with Oak Island Road. Turn Right.
0.7 km Bend to right on road marked Bluff Road.
2.0 km Cross railway tracks.
2.3 km Turn left on dirt road to beach.
2.4 km Beach. Turn right (southeast) and walk ~200 m to first area of outcrop.

Details

This stop is intended to show the undeformed characteristics of the Horton Bluff Formation, a unit that will be encountered in various deformed states at many of the later stops on the trip.

The Horton Bluff Formation is characterized by interbedded shales, sandstones, and impure dolomitic carbonates that probably represent paleosols. The section was described by Martel and Gibling (1991, 1996) and interpreted as a classic lacustrine succession, dominated by successive shallowing-upward cycles representing the filling of the lake following each subsidence episode. Although relatively undeformed (at least compared with later stops) the succession contains a number of soft-sediment features. Most notably, synsedimentary dykes are common, and frequently feed upward into conspicuously thickened, lenticular units of overlying sandstone. These were interpreted by Hesse and Reading (1978) as feeders overlain by sand volcanoes. However, Martel and Gibling (1993) reinterpreted the sand lenses as hummocky cross-strata formed by storm waves; the underlying sedimentary dykes were interpreted as products of transient pressures generated by waves impinging on a lake shore.

If the tide is low enough, we may be able to see a large, roughly elliptical region of disturbed bedding on the foreshore. This is also interpreted as some kind of dewatering structure, though its origin has not been entirely satisfactorily explained.

Retrace steps to highway 101 at Avonport and turn south towards Halifax

SPLIT ROCK: JOHNSON COVE TO MUTTON COVE

Access

Travelling south on highway 101 toward Halifax. At exit 4 turn off and take highway 215 east toward Newport (0 km).
3.8 km Newport Corner; - turn left just after the military communications antennae on the left.

9.3 km	*Brooklyn; turn right following highway 215 and 14. This is the first of a rather complex series of intersections over the next 2 km, through which we will steadfastly follow highway 215.*
10.0 km	*Bear left to follow highway 215 as highway 14 turns off right.*
11.0 km	*Turn left following 215*
11.2 km	*Bend right following 215*
13.2 km	*Kennetcook River Bridge. Note the outcrops of deformed Windsor Group on the south bank of the river*
20.5 km	*Centre Burlington Sandford's store*
28.0 km	*Avon Emporium*
36.4 km	*Road runs beside shore in Cheverie*
38.1 km	*Turn left on Sherman Lake Road*
39.0 km.	*Park at beach. Proceed on foot walking north along shore.*

Although a number of 'stops' are identified in the 3 km coastal section (Fig. 3-4) between Johnson Cove and Mutton Cove, exposure is nearly continuous after the first covered interval. Be prepared to see an enormous variety of structures, including many that cannot be described in this account.

Vehicles will return to highway 21 and proceed to the end of the section

39.9	*Turn left and proceed north 900 m to Ocean Beach Road*
40.8.	*Turn left on Ocean Beach Road and proceed to the beach.*
42.7	*Park in beach parking area at Mutton Cove.*

SR-1: Cheverie Formation

At the beach there are typical exposures of the Cheverie Formation (the upper portion of the Horton Group in this area, conformably overlying the Horton Bluff Formation). The Cheverie Formation consists of red to yellow sandstones and mudstones. The laminated sandstones locally include sedimentary structures such as cross-bedding and climbing-ripple cross-lamination. Mudstones locally display mudcracks, plant fragments, and yellowish carbonate concretions interpreted as caliche. Both sandstone and mudstone locally contain upright fossil tree stumps, though typically not particularly well preserved. The Cheverie Formation is interpreted as a fluvial succession probably deposited in channels and arid floodplains. There are significant differences between this succession and that on the SW side of Cheverie Harbour, suggesting significant lateral variability.

SR-2: Uppermost Cheverie - Basal Windsor section

The uppermost beds of the Cheverie Formation are calcareous sandstones. The transition to overlying pinkish grey fine limestones of the basal Windsor Group (Macumber Formation) is inconspicuous but sharp. The Macumber Formation consists of laterally continuous laminated limestone with a yellow to reddish orange oxidized weathering surface. The resistant limestone forms a distinctive linear outcrop on the shore (Fig. 3-4). The overall stratigraphic thickness of the Macumber Formation is roughly 2-3 metres, with individual laminations of 1-2 mm in thickness.

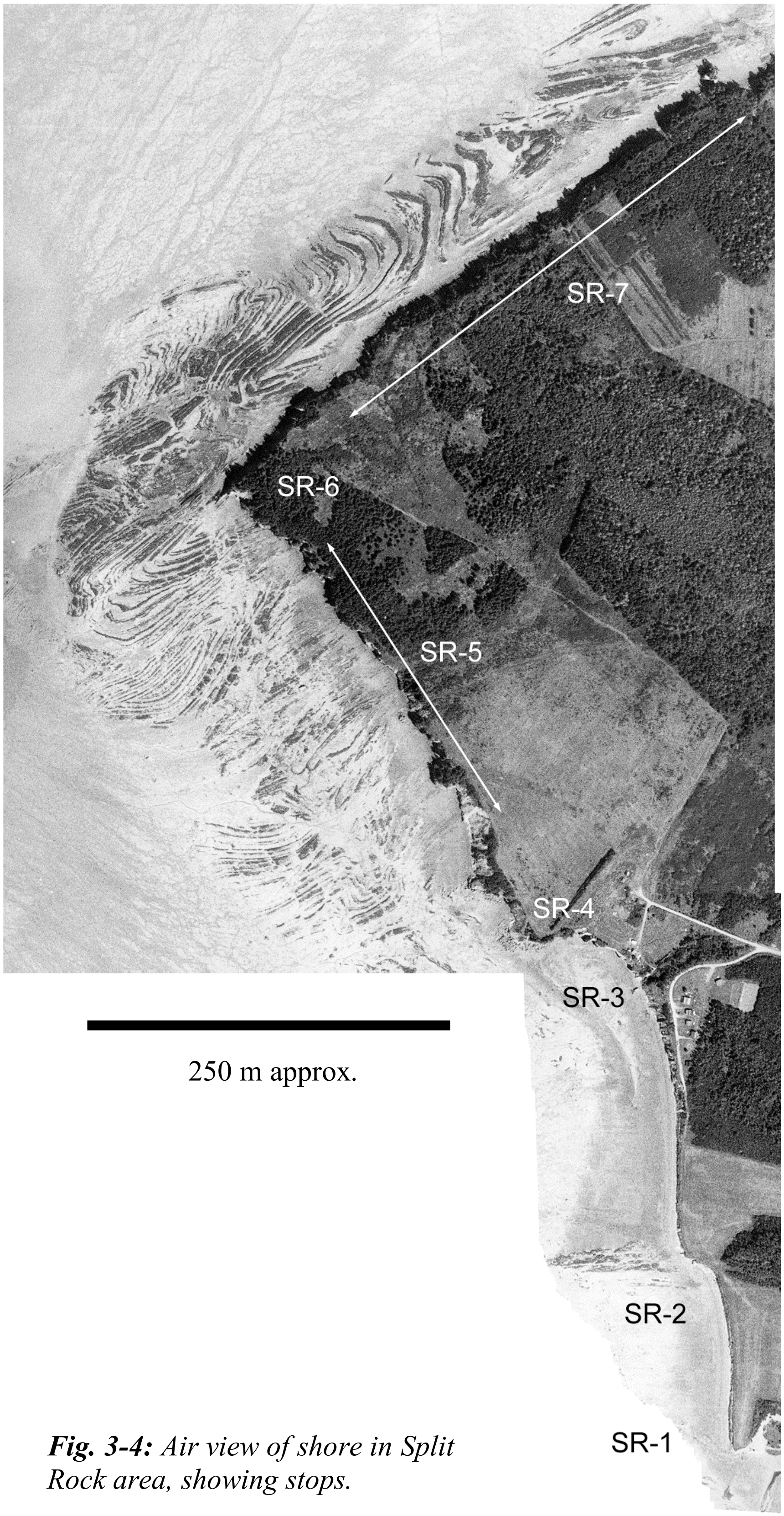


Fig. 3-4: *Air view of shore in Split Rock area, showing stops.*

Several different interpretations have been proposed for the Macumber Formation and associated facies at the base of the Windsor Group, summarized by Lavoie et al. (1995); interpretations include intertidal algal mats (Schenk 1967), deep water methane-vent bioaccumulations (von Bitter et al., 1992), saline lake deposits (Schenk et al. 1992), and deep water microbial mats (Lavoie et al. 1995).

The Macumber Formation passes transitionally upward into a poorly understood unit known as the Pembroke breccia. It consists of variably sized, randomly oriented, elongate angular blocks of thinly laminated yellow weathered limestone, of Macumber Formation origin (Lavoie et al. 1995), typically in a limestone matrix. The origin of the Pembroke breccia is controversial. Four different processes may have contributed: (1) synsedimentary slumping produced a breccia of well bedded microbial mats, and slump folds in contorted mats (Lavoie et al. 1995). (2) Because of the absence of certain stratigraphic units above the Macumber, Lavoie et al. (1995) also proposed that some of the breccia is tectonic, formed by bedding-parallel extensional shear. Stratigraphic omissions were linked to the proposed Ainsley detachment positioned at the Macumber / evaporite boundary (Giles & Lynch 1994, Lavoie & Sangster 1995). (3) A late karstic breccia can incorporate both intact Macumber Formation and previously brecciated material of origins 1 and 2 (Lavoie et al. 1995). (4) Parts of the Pembroke breccia can also be interpreted as solution collapse breccia.

At Johnson Cove only a short interval of Pembroke breccia is exposed. Locally, cavities within the breccia include soft red sandstone possibly derived by infiltration from younger Carboniferous or Triassic units above. To the north, there is only sporadic exposure in the beach, including both Windsor and Horton lithologies. Bedding orientations appear chaotic, suggesting that there is a significant thickness of megabreccia, involving material from both groups. This zone is interpreted as the base of "allochthonous" Horton Group.

SR-3: Horton Bluff Formation, and Triassic Fundy Group graben-fill

At the start of continuous exposure (just south of SR-3, Fig. 3-5) there are Horton Group shales and thin sandstones that are tightly folded, with incipient axial planar cleavage in some beds. Immediately north, the sedimentary rocks are cut by a diabase dyke, and by faults that bring down a small graben-like outcrop of the Triassic Fundy Group to beach level. The Fundy Group is much more friable than the adjacent Horton Group and is easily distinguished. A first interpretation might suggest that the intrusion is related to the graben formation. However, this seems not to be the case because the intrusion closely resembles others in the area that have been dated as mid-Carboniferous, and appears quite unlike the Mesozoic basalts of the Fundy Group. Hence we have to regard the juxtaposition of the dyke and the graben either as a coincidence, or as a result of the localization of Mesozoic faults at pre-existing intrusions.

SR-4: Mid-Carboniferous intrusions

An approximately homoclinal, gently dipping succession of Horton Bluff Formation is exposed on the south-facing cliff to the north of SR-3 (Fig. 3-4). Spectacular wave-ripple marks are visible on the rocky shore. Despite the apparent low degree of deformation, there is a steep cleavage in places, and intersection lineations are visible on bedding surfaces. The preservation of sedimentary structures suggests that the strain is low.

Close to the top of the cliff, a diabase sill is more resistant to weathering than most of the Horton Bluff lithologies. At stop SR-4, the sill is abruptly connected with a dyke, which cuts across stratigraphy before connecting with another sill-like segment, that disappears under the beach. Lenticular intrusions of diabase are present low in the cliffs, and suggest that the magma filled a series of generally en-echelon tension gashes, interconnected fractures, and bedding-plane cracks. This group of intrusions is one of several that occur within Horton Bluff Formation rocks in the area. Although the age relationships are unclear at this locality, about 2 km to the south, in Cheverie, Johnston (1999) was able to show in thin section that the intrusions clearly cross-cut the fabric of highly deformed shale (Fig. 3-5a), though they are themselves cut by faults. Kontak et al. (2000) used Ar-Ar methods to date another intrusion in this suite, located about 2 km inland, also cross-cutting structure in the Horton Group, at 315+/-4 Ma.

These relationships indicate that the Horton Group was deformed prior to ~315 Ma. The sediments of the Horton Group would have been less than 40 million years old at this time.

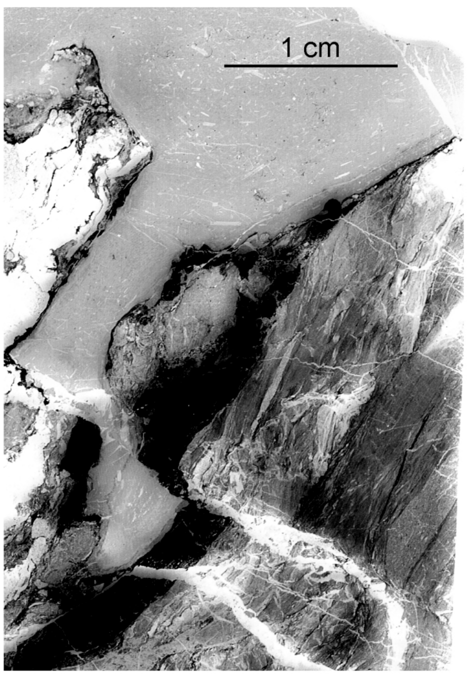


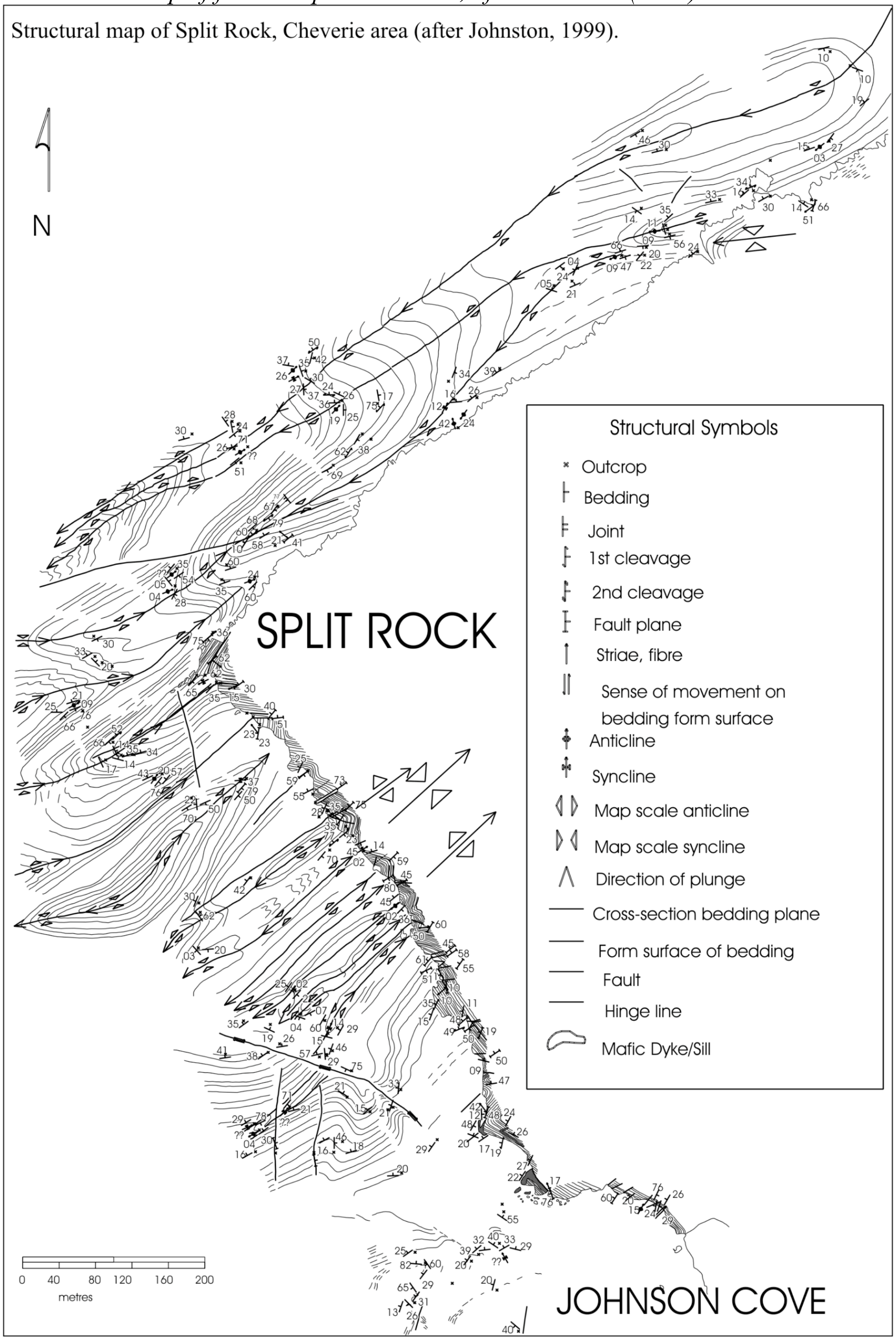
Fig. 3-5*: (a) Left: Thin section showing diabase intrusion (upper part of slide) cross-cutting fabric in shale (right). Both are cut by faults (lower part of slide). (b) Below: Syncline at Split Rock.*

SR-5: Folded Horton Bluff Formation

The coastal section to the north is unrivalled as a place to view folded sedimentary rocks. The folds are typically subhorizontal, upright to moderately inclined, with gently curved hinges that are locally spectacularly exposed in the beach (Fig. 3-4, 3-5b, 3-6). At some locations, folds are disharmonic, with shale intervals acting as detachment surfaces. Some folds are related to these detachments as detachment folds and fault-propagation folds. A spectacular example of the latter is exposed in the wave-cut platform.

Cleavage is weakly developed along this section of shore, though many bedding surfaces, especially in fissile shales, show a weakly developed crenulation lineation, parallel or sub-parallel with the main fold hinges.

***Fig. 3-6:** Detailed map of folds in Split Rock area, after Johnston (1999)*



Extensional structures are also much less conspicuous than those associated with shortening, but at numerous locations on this stretch of beach there are boudinage structures with axes either perpendicular to fold hinges or rotated slightly clockwise from this orientation.

Thin quartz and calcite veins mark some fractures, both parallel to layering and cross-cutting the bedding. Slickenfibres indicate the direction of slip. These have not been systematically investigated, but there appear to be veins that cross-cut the folds and veins that are folded, suggesting that several generations of fractures are present.

SR-6: Split Rock

The syncline at Split Rock is one of the best exposed on the shore (Fig. 3-4, 3-5b). This syncline has a near-horizontal hinge, but because of the slope of the beach, it is possible to view the fold in 'axial projection' from a suitable (if muddy) vantage point on the foreshore. A bed that is characterized by large 'cannonball' concretions of dolomitic carbonate can be located easily on both sides of the main fold. North of the fold hinge, bedding is vertical to overturned, and cleavage becomes more intense. A penetrative slaty cleavage is developed in black mudrocks; weak spaced cleavage occurs in some sandstones.

The dependency of buckle fold wavelength on layer thickness is well illustrated in muddy sections just north of the main syncline. Thick sandstone layers show long-wavelength buckles whereas thin layers are folded at a much smaller scale.

SR-7: Split Rock to Mutton Cove west side

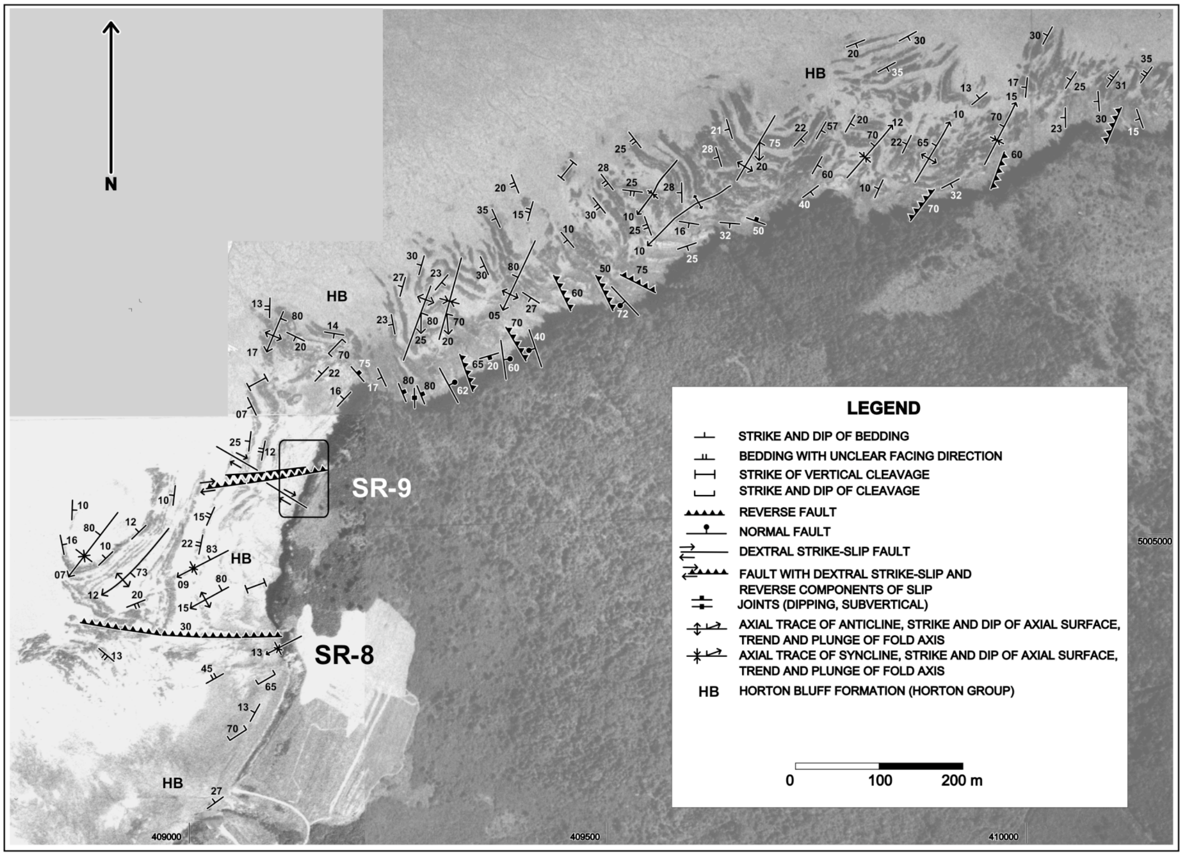
The hike along shore to Mutton cove (Fig. 3-4) is less spectacular than that to the south because the cliffs are roughly parallel to fold hinges, so folds are not seen in profile. However, the wave-cut platform displays multiple examples of folds that plunge both to the SW and to the NE. Mutton Cove itself is marked by a region of poor exposure that probably corresponds to a zone of NW-striking fractures. Examples of minor fractures in this orientation may be seen in the cliffs on either side of the cove. Mutton Cove makes an appropriate and accessible lunch stop.

SR-8: East side of Mutton Cove

At the resumption of outcrop on the east side of Mutton Cove (Fig. 3-7), folds are again prominent in the rocks of the wave-cut platform. Notice that the mudrocks are strongly cleaved, and there is a conspicuous, non-penetrative bedding-cleavage intersection lineation visible on the surfaces of sandstone beds. At several locations the cleavage is not parallel to the axial traces or hinges of the folds: - the folds are 'transected' by the cleavage in a counter-clockwise sense.

There are two 'classical' interpretations for this phenomenon. In the first, the folds and cleavage are regarded as separate generations of structures. Shortening occurred first in a more north-south direction, producing folds. Subsequently the shortening direction changed to a more east-west sense, overprinting the folds with cleavage. A second explanation involves dextral strike-slip or transpressional motion. Folds nucleated early, and underwent clockwise rotation with progressive deformation. Cleavage, on the other hand, reflects the bulk strain, acquired during the whole deformation history including components of shortening acquired up to the last moment of deformation. Cleavage therefore undergoes less rapid clockwise rotation, and ends up transecting the folds in a counterclockwise sense. It is not possible to choose absolutely between these alternatives based on the evidence at this spot, but the presence of major strike-slip features just to the NE (stop SR-9) favours the second explanation.

Fig. 3-7: *Aerial photo and geological map of the east side of Mutton Cove.*



SR-9: Strike-slip faults and positive flower structure

Several faults cut the beach to the east of Mutton Cove (Fig. 3-7, 3-8), and can be traced into the cliffs where they are seen to have steep dips (Fig. 3-9). Mineralized sheets with slickenfibres of quartz and calcite are found on the fault surfaces, and in some cases allow the sense of slip to be determined as predominantly dextral strike-slip. There do not appear to be any unambiguous piercing points that allow a clear evaluation of the amount of slip.

Folds are present in bedding adjacent to the largest, most continuous fault. The folds are oriented in an *en echelon* relationship to the fault, and die out into surrounding bedded sedimentary rocks. Cleavage is also present, oblique to the fault; cleavage planes curve to an orientation more nearly parallel to the fault in the immediately adjacent wall rocks.

Several fault-bounded slivers are located in the fault zone. One of these forms a prominent section of cliff, being 'popped up' relative to the surrounding rocks (Fig. 3-9c), and forming a structure closely resembling 'positive flower structures' often interpreted at much larger scale in seismic profiles from transpressional basins.

Fig. 3-8: *Detailed map of faults at stop SR-9, after Roselli (2004)*

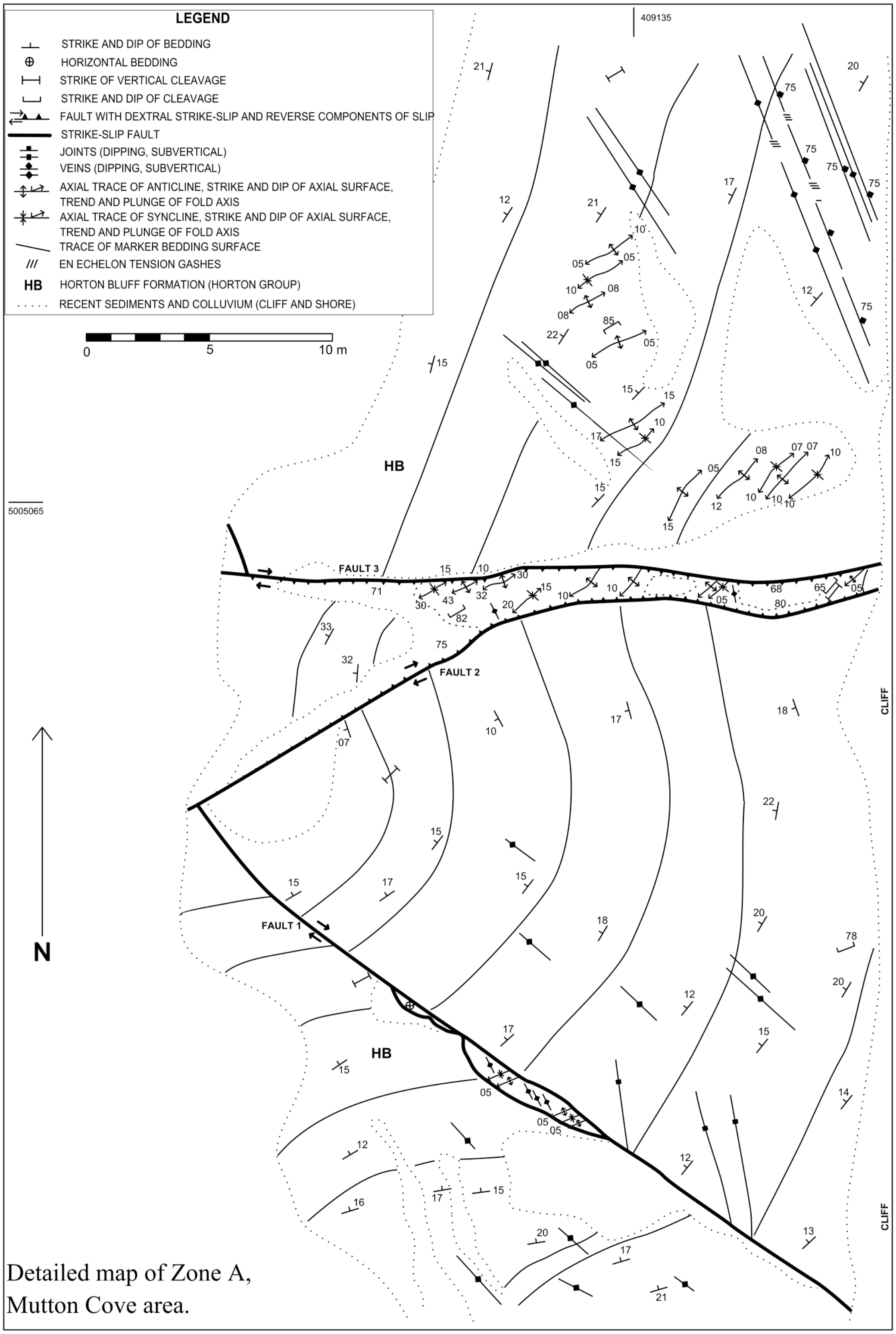
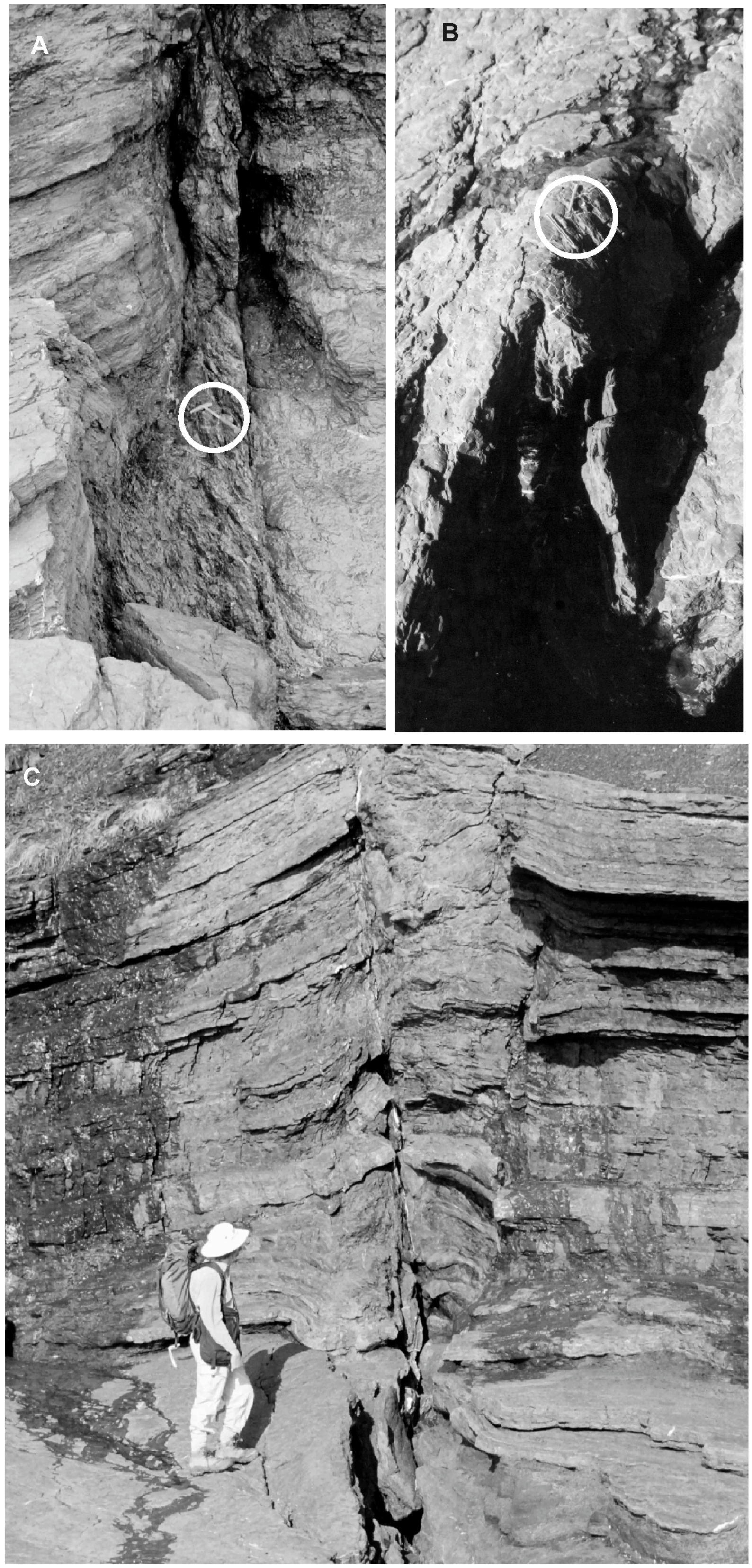


Fig. 3-9: *Photographs of structures on east side of Mutton Cove (Roselli 2004).*



Structures observed in the faulting zone, Zone A.

A) Restraining fault bends in Fault 1 contain folds and veins at the centimetre-scale (hammer for scale).

B)Transected cleavage observed in folds near flower structure (pencils for scale, parallel to fold axis and cleavage).

C)Traces of Faults 2 and 3 in the cliff mutually diverge, display strike-slip and reverse displacements. This fault zone is interpreted as a flower structure (person for scale).

RAINY COVE

Access

42.7 Leave Mutton cove and return to highway 215
44.6 Turn left on highway 215
47.3 Bass Creek
51.7 Cambridge Cove turnoff (golf course sign)
53.7 Rainy Cove - cross the bridge over Rainy Cove Brook and immediately park or turn onto dirt road to beach (Caution - there is no space to turn at the end of this track; it is necessary to reverse out most of the way. Large vehicles should remain on the highway.)

Introduction

The 400 m section at Rainy Cove is one of the best known in Nova Scotia, having been the subject of numerous student field trips from universities, high schools, and others. The section is particularly noted for spectacular exposures of downward facing folds (synformal anticlines and antiformal synclines) and for the angular unconformity between near-vertical Carboniferous rocks of the Horton Group and near-horizontal Triassic rocks of the Fundy Group.

Immediately west of the cove, the coastal section shows southwest-trending cliff-scale folds with weak axial-plane cleavage, somewhat similar to those at Cheverie, though the folds have been rotated into an orientation where their axial surfaces strike northwest-southeast. These folds overprint smaller, intrafolial folds that have very variable orientations, and strongly curvilinear hinges suggesting deformation while the sediments were still soft (Fig. 3-10). These are designated F1 folds, and the cliff-scale, cleavage-related folds are F2. The difference in orientation of F2 between this section and that at Split Rock suggests the existence of a third generation of folds F3, though these cannot be observed directly on the west side of Rainy Cove.

The traverse along the east side of the cove shows a progression from upward facing F2 structures to downward facing structures, demonstrating the refolding of inclined to F2 folds by a more upright generation designated F3. By careful mapping of the shoreline (Roselli 2004) it has been possible to disentangle these two generations of structures.

RC-1: Upward facing syncline and duplex structure

The southernmost portion of the Rainy Cove section is a mainly north-dipping cliff section interrupted by several faults that make correlation difficult. Depending on the state of erosion of the beach, excellent examples of deformed mudcracks, groove casts, and fossil trees may be seen in the cliffs and on the beach.

A broad northeast-plunging syncline with a general box-fold shape (D, D1 and D2 in Fig. 3-12 and Fig. 3-13) brings bedding to near horizontal. The north limb occupies 50 m of cliff section and is interrupted by a fault at the north end. The north limb of the fold is marked by repetition of bedding in a duplex of multiple horses and associated folds, indicative of significant shortening. Is this duplex an accommodation structure produced during folding or is it an earlier structure that has been refolded by fold D? The vergence of folds (Z-sense as viewed in the cliff) is inconsistent with the geometry of fold D, suggesting refolding.

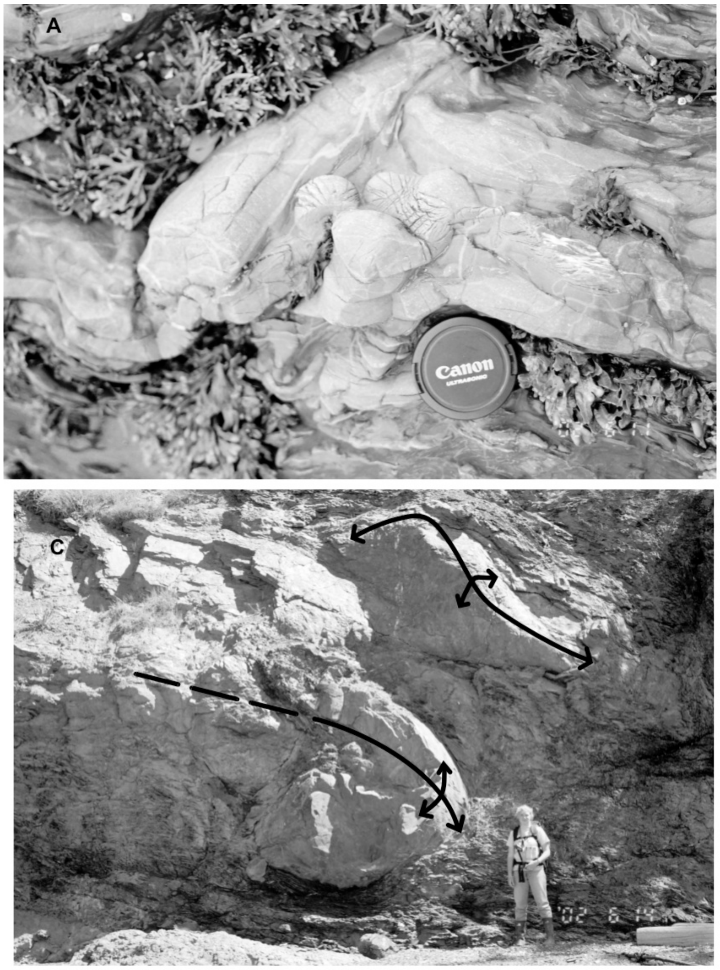
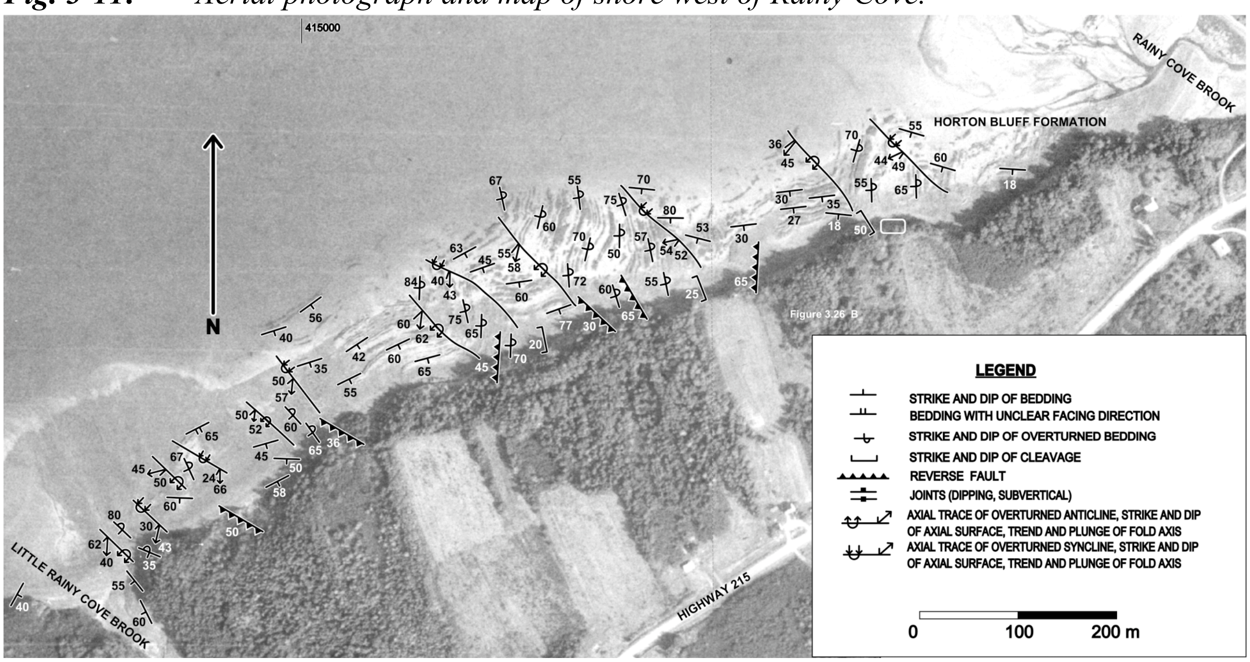


Fig. 3-10: *F1 folds and associated structures in the Little Rainy Cove area. (A) Convolute lamination and hydroplastic deformation structures in fine sandstone folded beds (lens cap for scale). (B) Variably oriented F1 folds in the cliff (person for scale).*

Fig. 3-11: *Aerial photograph and map of shore west of Rainy Cove.*



RC-2: Multiple upward facing folds

North of the fault, a section of cliff exposes near-continuous stratigraphy in a south-younging, steeply north-dipping overturned section. Towards the north end of this section, cascades of upward facing S-folds cut the strata. Locally there is a very weakly developed north-dipping cleavage that has a similar (not identical) orientation to the fold axial surfaces.

A conspicuous and complex anticlinal structure (Fold E in Fig. 3.28 and Fig. 3.31, Fig. 3.33 B) occurs in sandstone at the north end of this section. The central core of the fold is almost isoclinal, with tightly adpressed limbs. It is surrounded by a box-fold. Depending on the level of the beach, it may be possible to observe that a stratigraphically underlying shale section is much less tightly folded and much less shortened, demonstrating that the tight fold in the sandstone is a detachment fold.

RC-3: Downward facing folds: synformal anticline and antiformal syncline

About 20 m north of fold E is the first of a series of cliff-scale folds which face downward. The first is an anticline A whose hinge can be seen in the cliff, plunging moderately southwest in an axial surface that dips southwest. The fold is a reclined fold - one in which the hinge is aligned down the dip of the axial surface. The facing direction in the axial surface is very slightly downward, towards the southeast.

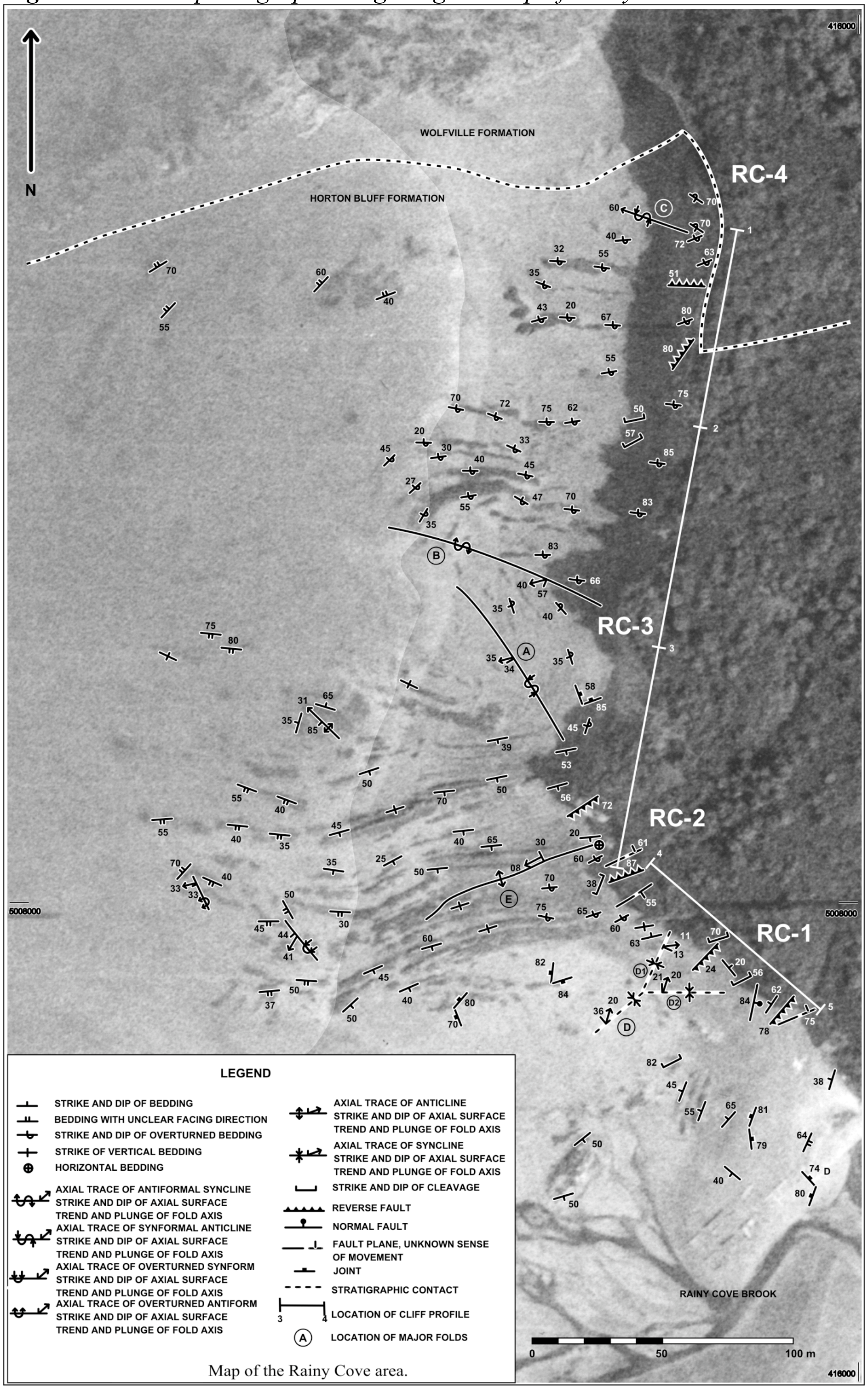
In the cliff, it is possible to see the folded under-surface of a sand bed bearing load casts, cut by a number of parasitic buckle folds and very weak crenulation lineations suggesting that this fold is related to the weak cleavage seen elsewhere in the shore. This fold is interpreted as an F2 fold.

About 40 metres to the north is a second large downward-facing fold, antiformal syncline B. This fold plunges moderately west. The trace of folded bedding is picked out by seaweed-covered sand ridges on the foreshore. The axial surface is moderately inclined, dipping to the south-southwest. Facing direction is clearly steeply downward, into the cliff to the east, as indicated by cross-laminations on the north limb, and also by tracing the way-up from fold A.

Caution: because of the seaward-dipping bedding surfaces in the hinge region, this fold has been the site of numerous cliff-falls in the past. (The photograph Fig. 3-14b shows its state approximately 20 years ago.) Beware of falling rocks. Do not stand close to the cliff and be especially aware of overhangs.

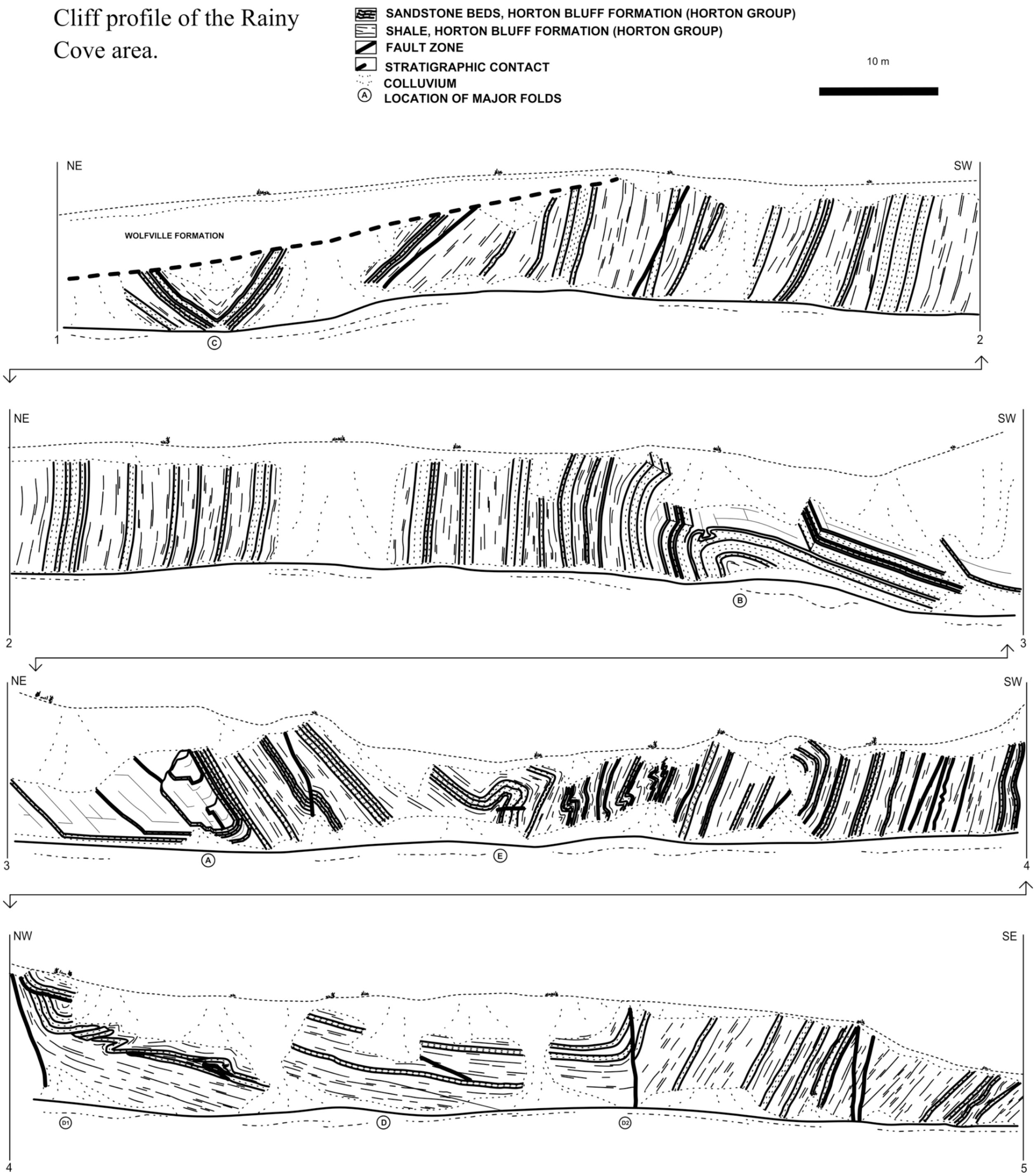
The axial traces of folds A and B converge offshore, raising the question of what happens when they meet; do the two folds merge or does one refold the other? Unfortunately the sand ridges are lost in the mud before this question can be resolved. However, based on cleavage orientations measured at intervals along the cliff, it seems likely that fold A is earlier, and is refolded by fold B.

Fig. 3-12: *Aerial photograph and geological map of Rainy Cove shore*



Map of the Rainy Cove area.

Fig. 3-13: *Sketch cliff profile at Rainy Cove: locations shown in Fig. 3-11.*



RC-4: Unconformity and Fundy Group

Caution: the remaining sections of the cliff are steep and prone to rock-falls. View the structures from a distance and in the wave-cut platform. Do not stand under the overhanging or vertical sections of the cliff!

North of fold B, there is a long (~100 m) homoclinal section of cliff with near-vertical bedding, younging south. Toward the top of the cliff, strata of the Triassic Wolfville Formation can be seen, dipping gently north, and resting with profound angular unconformity on the Horton Group below.

Several small folds can be seen both in the cliff and in the foreshore. In general, these cannot clearly be related to the large downward facing folds B and C; they are believed to pre-date the larger structures.

Fold C is a synform located in the cliff immediately below the Mesozoic unconformity. Tracing way-up from nearby accessible outcrops with cross-laminations shows that this fold is an anticline, a third major downward-facing fold in the profile. A smaller asymmetric fold on its south limb, with weak axial plane cleavage, has the wrong asymmetry to be a parasitic fold on fold C. It probably belongs to an earlier generation.

Interpretation

The main challenge in interpreting the Rainy Cove section is the interpretation of the downward facing folds A, B and C. In general, downward facing folds cannot be easily produced by a straightforward shortening of originally upright, horizontal stratigraphy; some kind of fold overprinting is required, either to invert stratigraphy before later folding, or to rotate already-formed folds into a downward-facing orientation.

Two geometries can explain the configuration of folds at Rainy Cove. In the first (Fig. 3-15a), folds A, B, and C were produced in an early episode of recumbent, south facing F2 folds, and these have been overprinted by upright F3 folds D and E, bringing them into their current downward facing orientation. In hypothesis 2 (Fig 3-15b), only fold A is a recumbent F2 fold. Folds B, C, D and E are all later, superimposed upright F3 folds; B and C end up with downward facing directions because they are overprinted on the overturned limb of fold A.

Scattered observations of cleavage along the section at Rainy Cove are roughly parallel to the folded axial surface of early fold A in hypothesis 2. In contrast, the cleavage observations, particularly the north-dipping cleavages observed near fold D, are not parallel to either set of axial surfaces in hypothesis 1. Hence we favour hypothesis 2: the downward facing folds B and C are developed in the overturned limb of a large recumbent anticline, the hinge of which is exposed as fold A.

Fig. 3-14: *Downward facing folds A, B and C at Rainy Cove.*

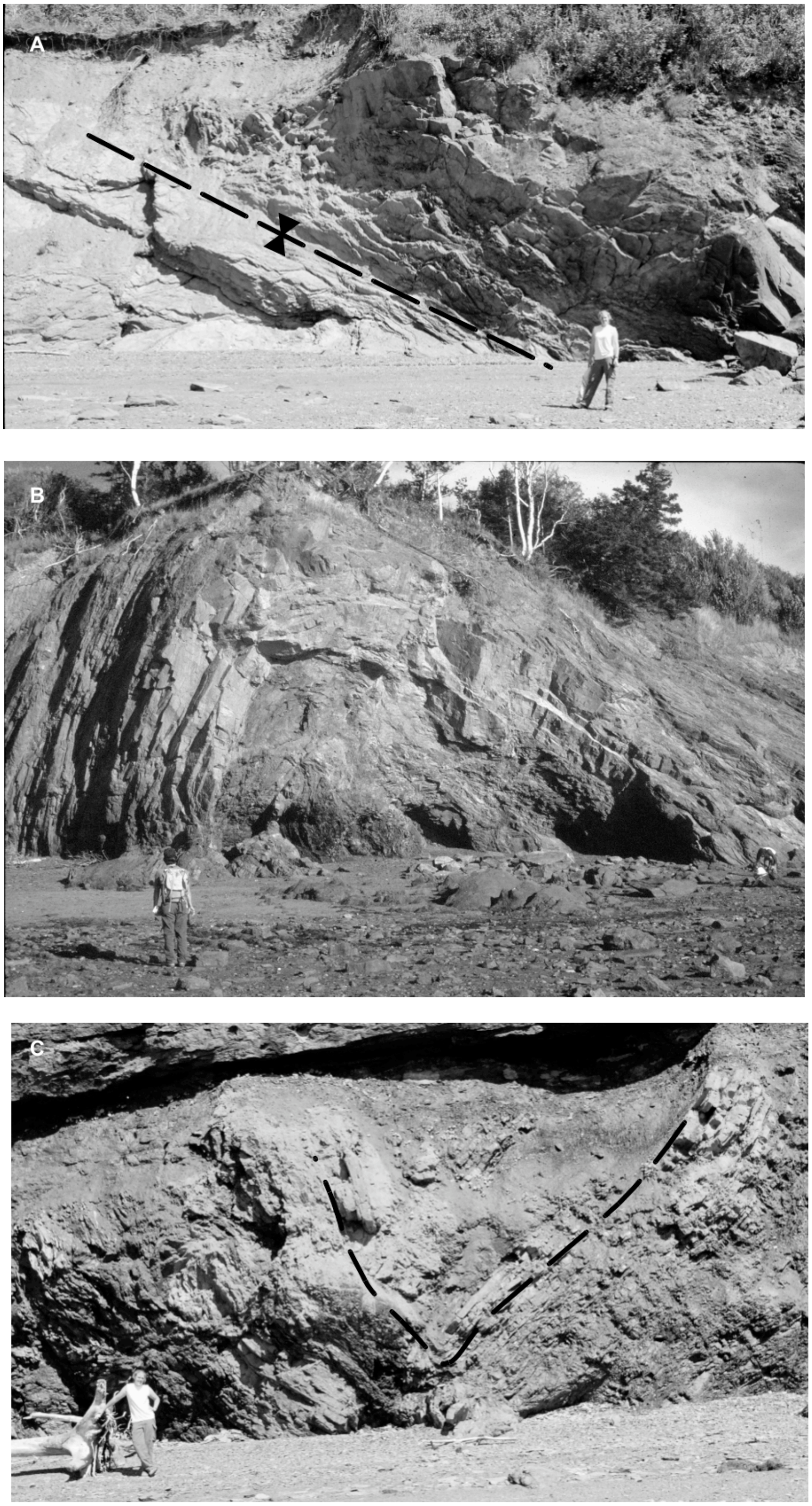
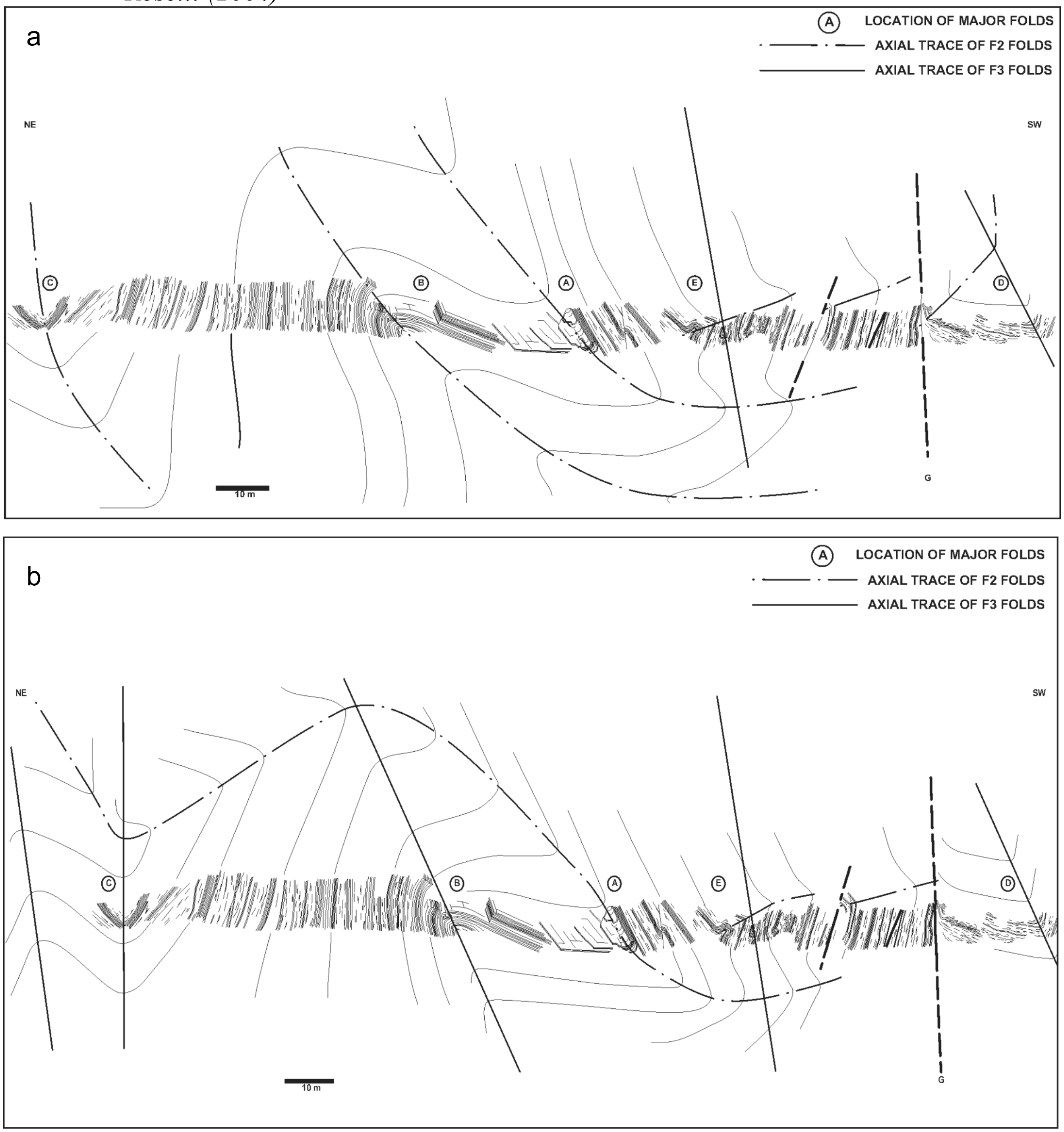


***Fig. 3-15**: Alternative interpretations of the cliff profile at Rainy Cove (see Fig. 3-13). After Roselli (2004)*

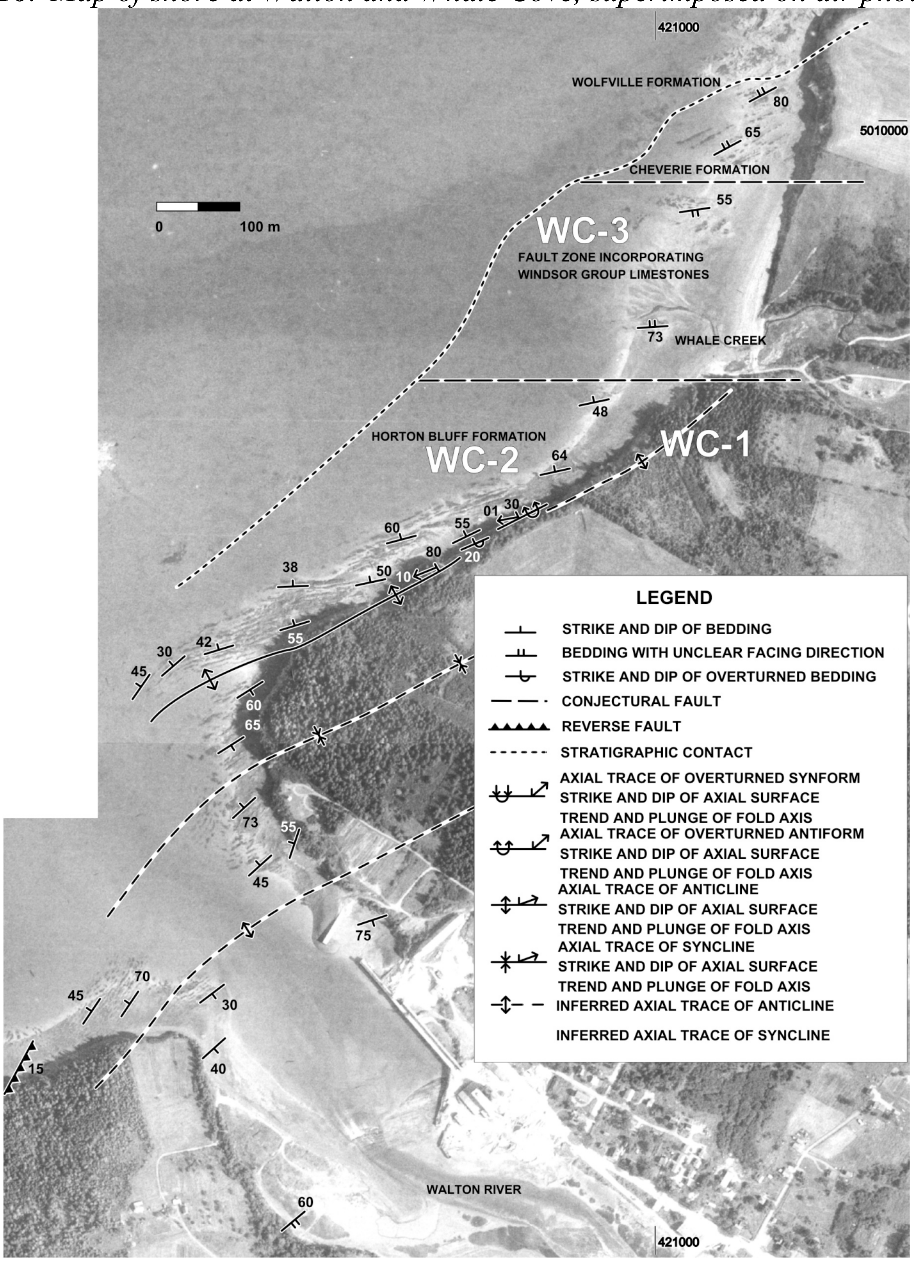


WHALE COVE

Access

53.7 Leave Rainy Cove heading east on highway 215.
69.0 Intersection at centre of Walton community. Turn left (though still following highway 215)
70.7 Stop at Whale Creek campground entrance and walk to the shore. At the shore turn left and proceed about 100 m along the beach (Fig. 3-16).

Fig. 3-16*: Map of shore at Walton and Whale Cove, superimposed on air photo.*



WC-1: Amphibian footprints

Notice the fine set of fossil amphibian footprints in rippled siltstone of the Horton Bluff Formation.

WC-2: Refolded folds

The cliff at this location shows two generations of folds (Fig. 3-17). A recumbent set of tight folds (tentatively identified as F2) is refolded by an upright antiform (probably F3). This section is a small-scale version of the structure seen at Rainy Cove, and illustrates how downward-facing folds originate by the refolding of a large-scale recumbent fold. At the foot of the cliff are small-scale examples of east-west tension gash zones with en echelon gashes indicating dextral shear.

Fig. 3-17: *Structures at Whale Cove*

WC-3: Fault zone in Horton and Windsor Groups

If time and tide permit, cross the creek to observe a large scale zone of disrupted stratification and foliated fault rocks that strike approximately east-west through the cove. Rocks in the zone include foliated fragments of Windsor limestone and dolostone. This is one of the largest fault zones in the area; frustratingly it is exposed over only a short distance; to the north and west, the Horton Group disappears under cover of Triassic Fundy Group rocks.

Travel to Truro

70.7 km	*Leave Whale Creek campground, heading east.*
113.7	*Maitland, Capt. Douglas House Inn.*
121.4	*Intersection; highway 215 joins 236. Go straight on*
121.6	*Turn left on highway 236*
123.1	*Bridge over Shubenacadie River*
124.6	*Intersection; take straight on road which is highway 289 signposted to Brookfield*
137.4	*Take on-ramp to join highway 102 north towards Truro*
149.1	*Exit 13 take at off-ramp to Truro*
149.5	*Turn right into Truro on Truro Heights Rd.*

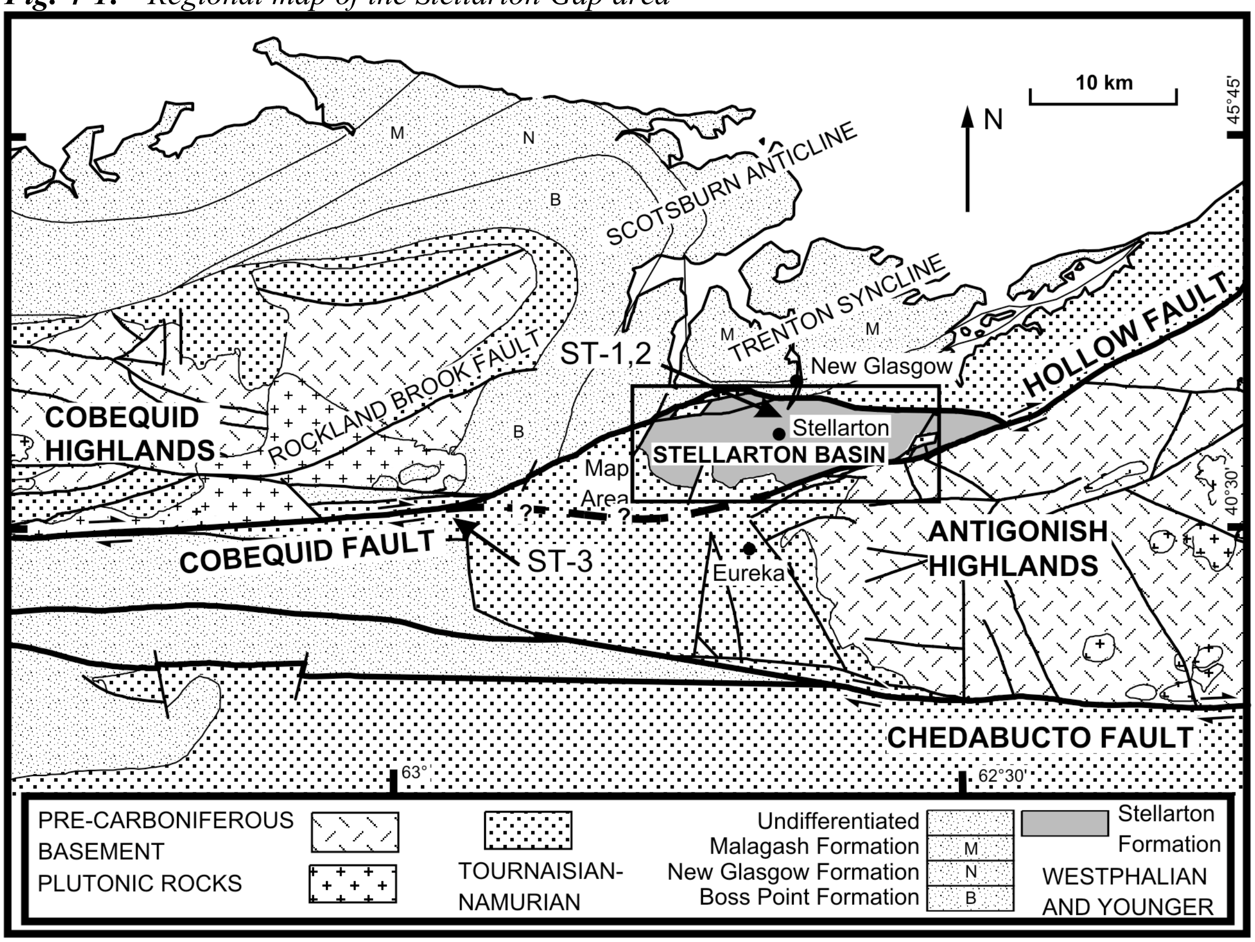
4. COBEQUID FAULT ZONE TRURO-STELLARTON

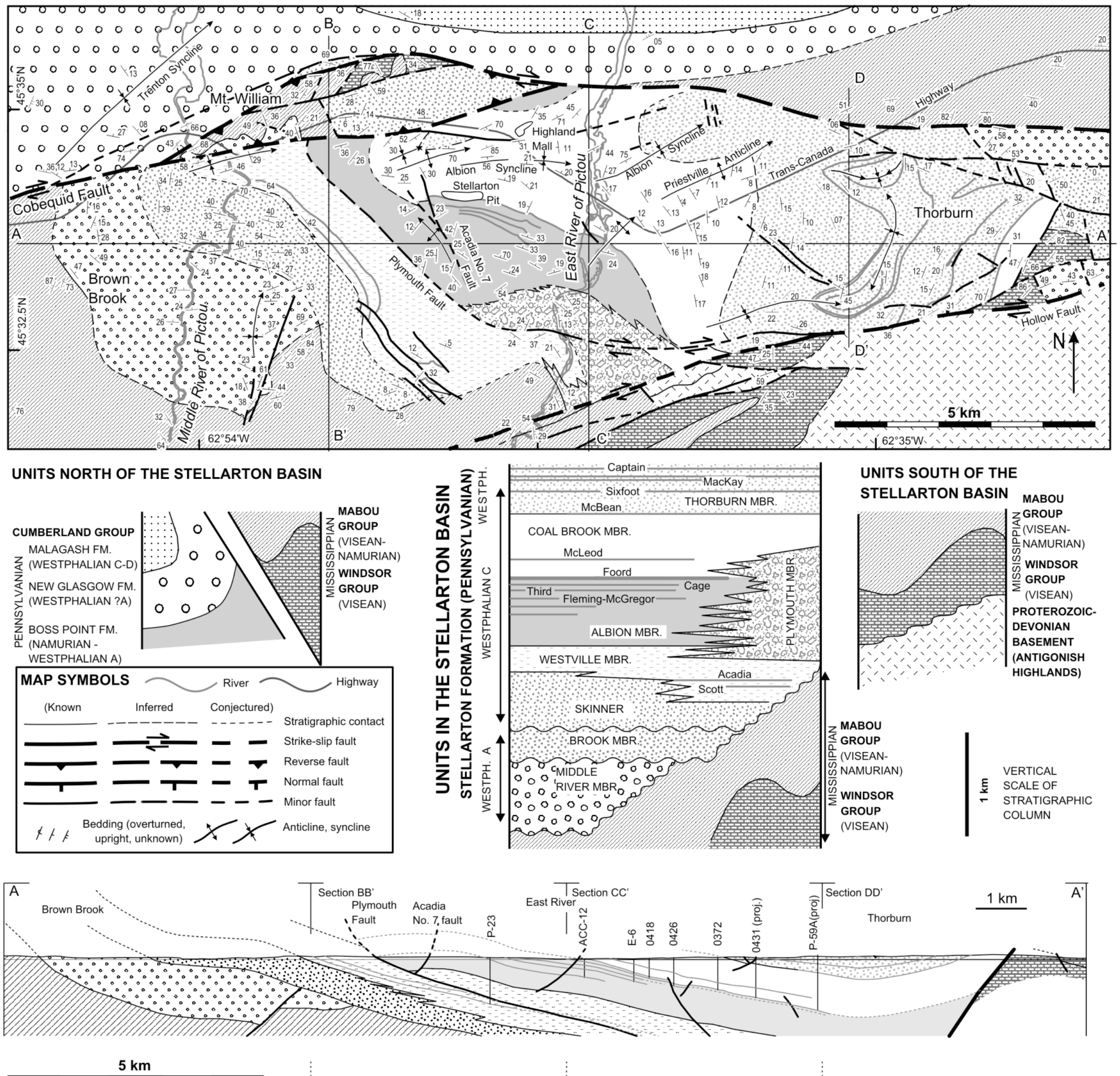
GENERAL GEOLOGY

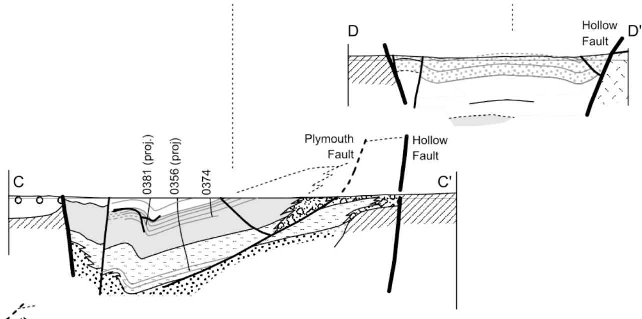
East of Truro the Minas Fault Zone extends along the southern margin of the Cobequids. The most conspicuous mapped lineament is the Cobequid Fault itself, which separates well indurated basement rocks of the Cobequids from softer, late Carboniferous to Triassic rocks to the south.

East of the eastern extremity of the Cobequids is a region in which Avalonian basement rocks are unexposed; Maritimes Basin sediments of Carboniferous age extend from the Minas Fault Zone northward as far as the Northumberland Strait, and beyond beneath the Gulf of St. Lawrence and Prince Edward Island. Farther west, Avalonian basement reappears in the Antigonish Highlands. The area between the Cobequid and Antigonish Highlands is known as the Stellarton Gap (Fig. 4-1). The Carboniferous (Westphalian) Stellarton Basin was formed in the central portion of this area.

Fig. 4-1: *Regional map of the Stellarton Gap area*







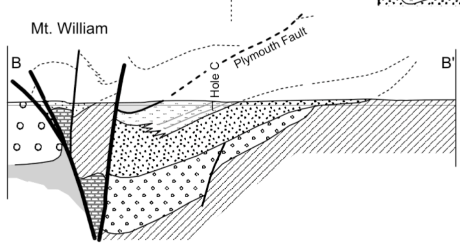
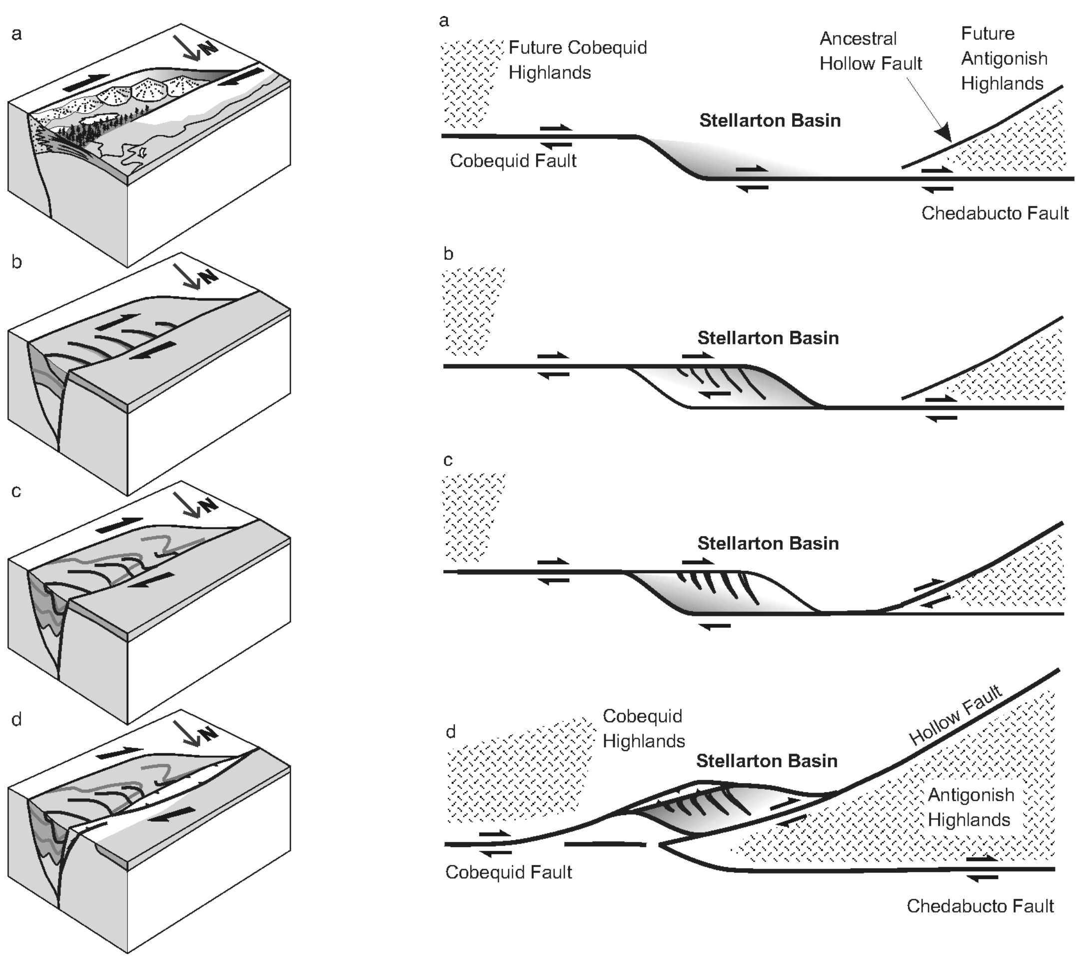


***Fig. 4-2**: Map and cross-sections of Stellarton Basin. Location: See Fig. 4-1*

The Minas Fault Zone splays into the Stellarton Gap. The Cobequid Fault curves northward towards New Glasgow. A deep basin, the Stellarton Basin, filled with Pennsylvanian Cumberland Group (Stellarton Formation), joins the Cobequid fault to the Hollow Fault, which runs along the NW margin of the Antigonish Highlands. The Stellarton Basin is interpreted as a transtensional basin, occupying a pull-apart structure developed at a releasing bend on the Cobequid-Hollow system. The basin is filled with about 3 km of Pennsylvanian sediments, mainly of Westphalian C age, including the entire Pictou coalfield. The Foord seam is up to 13 m thick, and is reputed to be the thickest coal seam in eastern North America.

Because of its history of coal mining and exploration, the subsurface structure of the Stellarton Basin is exceptionally well documented. The map and cross sections (Fig 4-2) show an extensional basin overprinted at its northwest edge by a positive flower structure, representing late-stage shortening. Based on variations in the thickness of its fill, the basin is interpreted to have subsided asymmetrically (Fig 4-3). Subsequent (latest Carboniferous or Permian?) interaction with irregularities in the MFZ associated with the east edge of the Antigonish Highlands was probably responsible for the overprinted compressional structures.

Fig. 4-3: *Interpretive diagrams showing evolution of Stellarton Basin. Left: Development of an asymmetric pull-apart basin subsequently subjected to transpression. Right: Possible regional location of Stellarton Basin during its development.*



FIELD STOPS IN THE STELLARTON AREA

Travel Truro to Stellarton

0.0	*Turn off Truro Heights Road to join highway 102, heading north.*
2.0	*Highway crosses Robie St. exit 14 bridge.*
3.9	*Highway divides - take ramp for highway 104 East.*
4.9	*Ramp joins 104 eastbound.*
36.7	*Highway crosses deep roadcut outcrop (stop ST-3).*
60.9	*Take exit 23 off-ramp.*
61.5	*Crossroads at foot of exit 23 ramp; go straight on MacGregor Ave.*
62.5	*Cross Foster Ave; go straight on MacGregor.*
62.7	*Entrance to Pioneer Coal Stellarton Pit.*

ST-1: Stellarton coal mining operations

Most of the Stellarton Formation is recessive and outcrops are scarce. The location of field stops will be determined by recent activities in surface coal-mining operations at the various pits active in the coalfield. Details will be distributed on the trip.

ST-2: Highland Mall, Stellarton

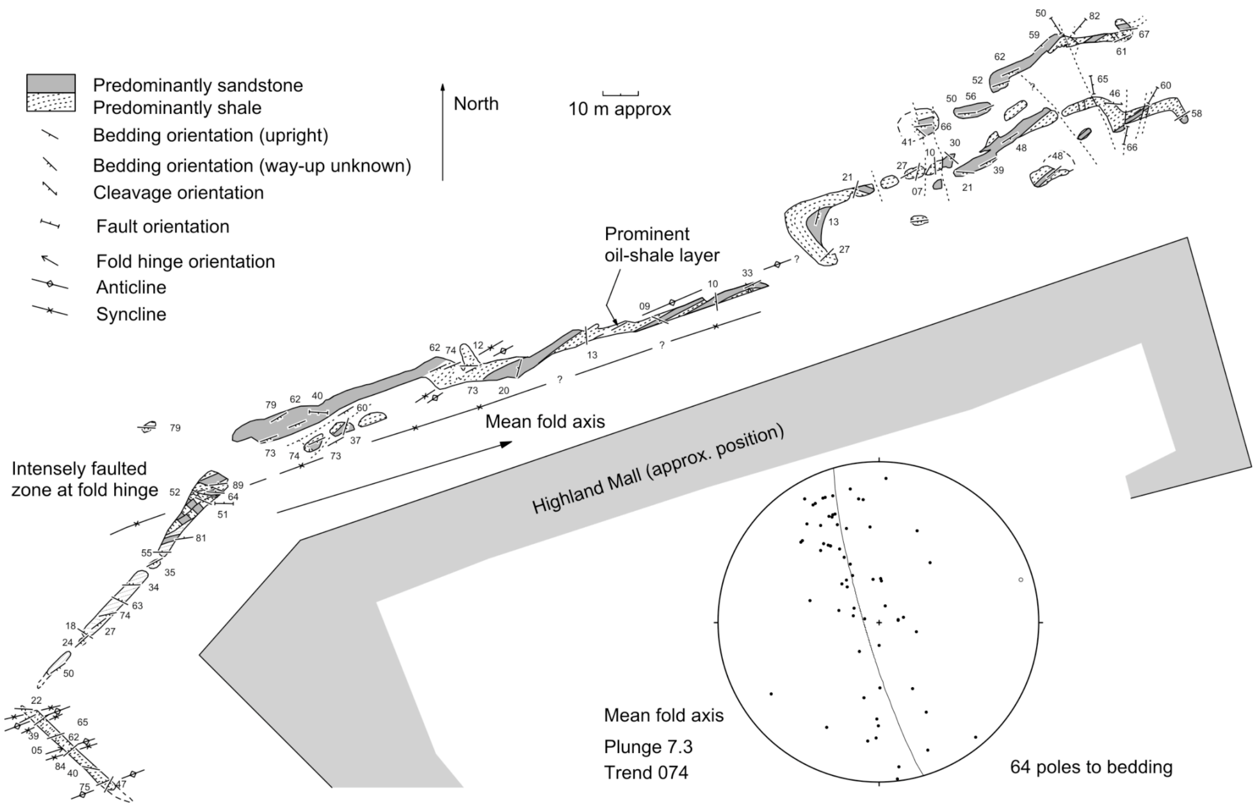
62.7	*From Stellarton Pit, head north on MacGregor Ave.*
64.1	*Crossroads near exit 23; turn right.*
64.3	*Pass under highway 104.*
64.5	*Turn left into Highland Mall; drive to far end of mall and park; proceed on foot to the long outcrop behind the mall (Fig. 4-4).*

This series of outcrops is representative of a block separated from the main section of basin fill by the Bridge Fault, but lying south of the Fletcher fault. The Bridge Fault lies immediately south, passing through exit 23 where you may have turned off the Trans-Canada Highway. The rocks in this slice can be tentatively correlated with those to the south of the Bridge Fault in the main part of the basin. To the north of this location, a historic mine, known as the Haliburton pit, was sunk into a south-dipping seam identified as the Foord, which marks the top of the Albion Member. The section in these outcrops is probably therefore correlative with the Coal Brook Member of the Stellarton Formation.

The outcrops (Fig. 4-4) appear complex and confusing at first, but will illustrate the characteristics of the north basin margin, which are also known somewhat from the plans of historic mines beneath this area. The overall structure appears to be a syncline whose axial trace almost parallels the outcrop face. This is cut by numerous north-south faults, most of which are extensional, but which show a variety of mineral fibre orientations that are consistent with a combination of dip-slip and strike-slip motion (present-day orientation). Lithologically, the rocks are shales and sandstones, locally with abundant fossil wood and fish scales, indicating fresh-water, lacustrine to fluvial environments.

At the east end of the outcrop, a south-dipping surface shows numerous extensional faults with a variety of slickenside striations. Note how the throw on individual faults varies rapidly along strike, producing 'scissors fault' geometry locally. Note also that mudcracks (?) outline a deformed pattern, indicating penetrative north-south shortening.

Fig. 4-4: *Pace and compass map of outcrop at Highland Mall, Stellarton*



About half way along the outcrop an intensely faulted and overturned section marks the axial trace of the major syncline that dominates the structure. To the southwest, the outcrops show predominant north dip; the few south-dipping bedding measurements show continuity (bedding passes through the vertical) with adjacent north-dipping strata. This indicates that the major fold is inclined, with its axial surface dipping north.

ST-3: Fault in Minas Fault Zone

65.8 Leave Highland Mall, turn right
65.9 Turn right onto highway 104 towards Truro
83.4 Take off-ramp at Exit 19
83.6 Turn right at end of off-ramp
83.8 Intersection with old highway 4; turn left
91.1 Turn left at sign to Watervale
92.1 Turn left again at sign to Watervale
92.8 Start of outcrop in road cut. Park at top of hill.

Caution: although this road carries little traffic, occasional fast truck traffic uses the road. Trucks tend to travel fast under the bridge. Keep off the roadbed and keep eyes and ears open for traffic at this locality

At Mount Thom a broad zone of fault breccia is developed within Mabou Group rocks (Namurian) and is exposed in a road cut beneath the Trans-Canada Highway. The fault apparently cuts Mabou Group both to the north and south. It is probably therefore a southern splay of the main Cobequid fault.

The fault zone includes highly folded and brecciated lithologies of the Mabou Group. In addition, just north of the highway bridge, there are mudrocks that appear only weakly deformed. In places these contain cubic casts of a mineral that has since dissolved, possibly halite. These mudrocks are tentatively interpreted as a residue resulting from the solution of evaporites that were formerly present in the fault zone.

92.8 km *Retrace route to Exit 19.*
102.9 *Rejoin highway 102 heading west toward Truro*
142.7 *Take off-ramp to highway 102 to if returning to Truro or Halifax; continue on highway 104 towards Parrsboro for remainder of trip*

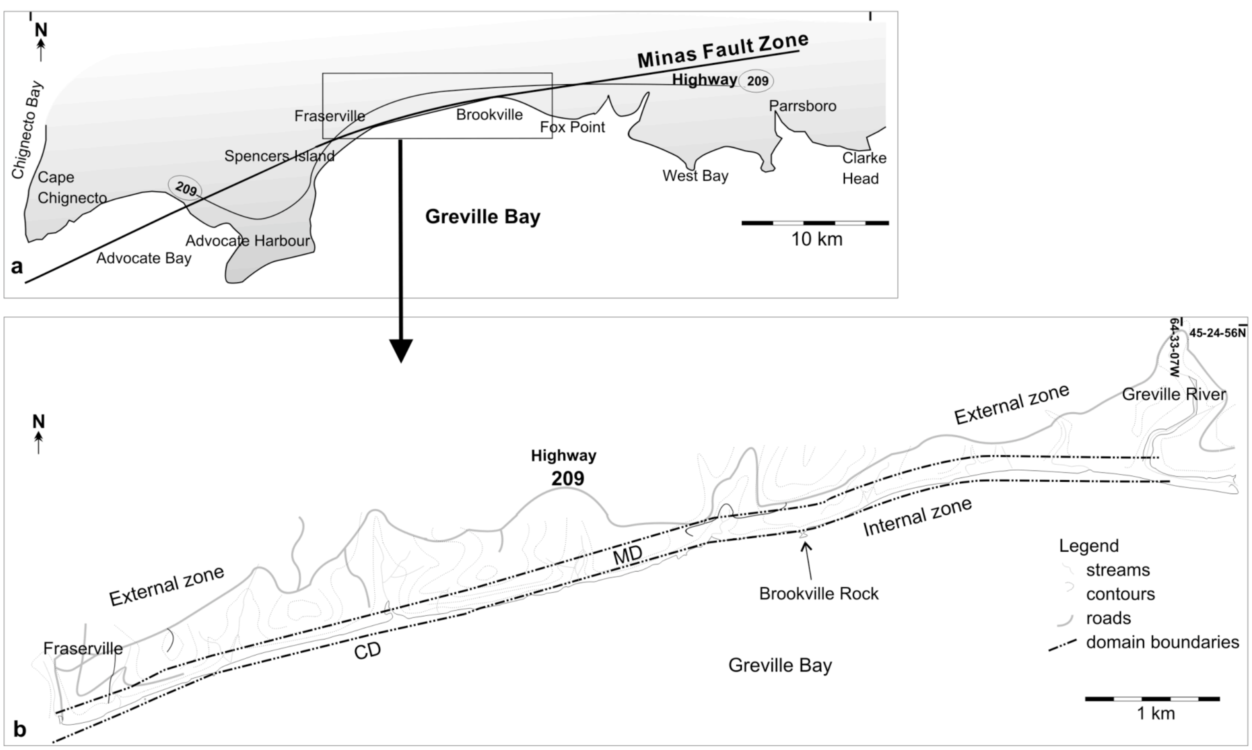
5. MINAS BASIN: NORTH SHORE

OVERVIEW OF THE GEOLOGY

Synthesis and Interpretation of the Minas Fault Zone

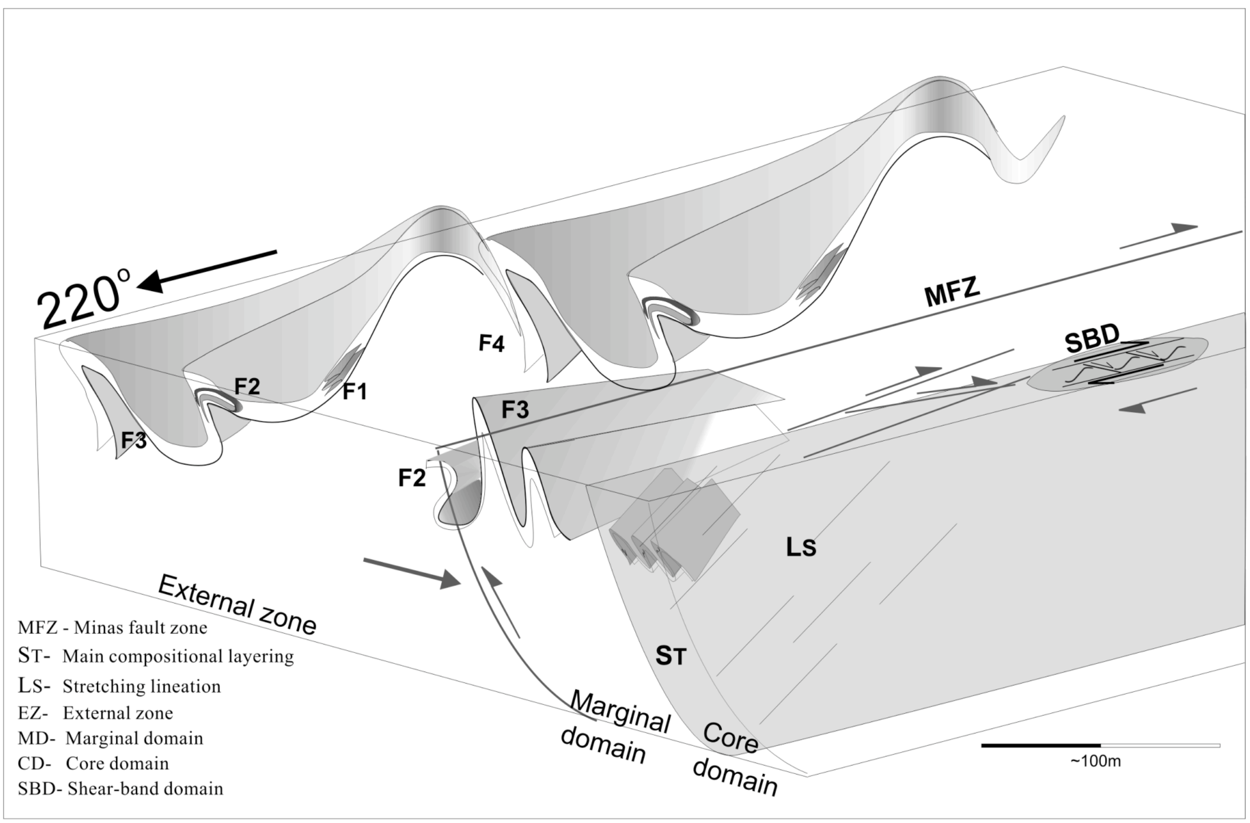
The Cobequid fault segment of the Minas Fault Zone extends approximately 60 km from Cape Chignecto along the north shore of Minas Basin before striking inland. The current Cobequid fault scarp delineates a moderately to steeply SSE-dipping surface that exhibits a swing in strike from 068º in the west to 082º in the east over a strike length of ~ 58 km (Fig. 5-1). The fault scarp separates zones of relatively low-strain (External Zone) and high-strain (Internal Zone), which are developed, respectively, NNW and SSE of this topographic boundary. The demarcation between the contrasting style and intensity of deformation of the external and internal zones is sharp. Both zones exhibit similar and related kinematic histories (MacInnes & White 2004), but the significantly lower levels of finite strain in the external zone enable fuller preservation of the sequence of deformation. The development of a first-order bimodal architecture comprising high-strain internal and external contractional zones is notably similar to models for transpressional zones (Sanderson & Marchini 1984, Jones & Tanner 1995, Lin et al. 1998).

Fig. 5-1 (a) Map of the Parrsboro area (b) Simplified map of Greville Bay showing the external and internal zones. The internal zone is divided into the marginal (MD) and core (CD) domains. The shear-band domain (SBD)occurs as lenses within the core domain.



Along Greville Bay, a segment of the Cobequid fault zone within the MFZ is exposed over a width of about 180 m south of the present fault scarp, trending approximately east-west. The southern boundary of the zone lies within the Minas Basin. The formations present in this coastal segment encompass Late Carboniferous Parrsboro Formation (Cumberland Group); Early Carboniferous Greville River Formation (Horton Group); Devonian-Carboniferous Fountain Lake Group rhyolite; and Neoproterozic Jeffers Group metasedimentary and metavolcanic units. Metamorphic grade is very low with apparent variations in metamorphic grade resulting almost solely from variations in finite strain. All the rock types have been incorporated into the fault zone as lithologically contrasting blocks separated by and containing breccia and gouge zones. Despite the geometric heterogeneity of the deformation, displacement sense, and vorticity indicators (fold vergence, S/C fabrics, shear-band displacements, Riedel fault patterns) consistently indicate dextral movement (pre-Mesozoic) throughout the MFZ.

Fig. 5-2: *Synoptic block diagram of deformation in MFZ. Overprinting relationships are easily identified in the lower strain External Zone. In the Internal Zone, layers are transposed and multiple generations of intrafolial folds can be identified. Ductile fabrics are overprinted by shear-bands (SBD) and Riedel fault arrays. View looking north.*



The Visean Windsor Group was incorporated into the fault zone during post-Windosr normal faulting. Fold structures in the Windsor Group fault slice trend parallel to the fault zone. They are thought to result from fault zone displacements. Overprinting the folds are dextral strike-slip faults, which accommodate displacement parallel to the trend of the MFZ (Fig 5-2). The Visean Windsor Group carbonates acted as detachment surfaces enabling horizontal displacement of overlying stratigraphic units. The Windsor Group fault slice, located at the contact between the marginal domain and the core domain, may have acted as an easy slip surface between the two domains. The Upper Carboniferous Parrsboro Formation was incorporated into the fault zone during post-Westphalian extension. The Parrsboro Formation was folded after incorporation into the fault zone. The contact between the Parrsboro Formation and the Horton Group is a dextral strike-slip fault; this indicates that dextral strike-slip displacement continued into post-Westphalian times. Regional studies suggest NW-SE-directed extension (Withjack et al. 1995), during the Permian(?) and Triassic, as Pangea started to fragment. Fold orientations in the Triassic Wolfville Formation indicate dextral strike-slip displacement in post-Triassic times. The Jurassic was dominated by large-scale extension during the complete break-up of Pangea and the formation of the present day Atlantic Ocean.

External Zone (not seen on field trip; similar to southern Minas shore)

The external zone (best exposed along Chigencto shore) is characterized by the development of structures typically associated with crustal contraction; these include: bedding-plane thrusts, ramp and flat structures, folds and duplex structures. Although the proximity of the Cobequid fault segment of the MFZ has induced higher strains than observed on the southern shore of Minas Basin, the early formed structures in both areas are remarkably similar. In general, early deformation comprises subhorizontal, bedding parallel deformation in both External and Internal Zones (Fig. 5-3).

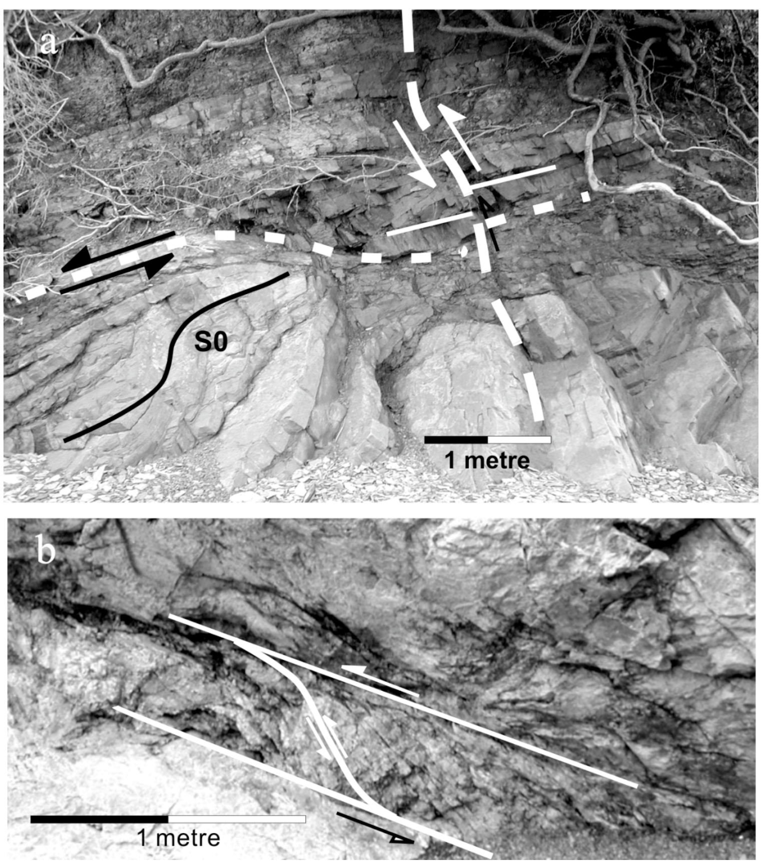


***Fig. 5-3**: Bedding-parallel thrusts typical of early transpressional deformation. (a) Low angle bedding-plane thrust faults in Parrsboro Formation are overprinted by high-angle reverse faults, both accommodating NNW-directed displacement. Photo facing N. (b) Thrusts within Windsor Group carbonate fault slice accommodate WNW-directed contraction. Greville Bay, internal zone. Photo facing E.*

Four generations of folds (Fig. 5-2) have been identified in the External Zone and record progression in the overall tectonic history. First generation folds ($F1_{EZ}$) deform bedding (S0) and a bedding parallel foliation, defined by dissolution-seams (Ss). These earliest recognized folds are tight to isoclinal with attenuated limbs. Isolated fold closures, typically of quartz-rich beds (S0), are preserved within later fold structures. First generation folds form in conjunction with bedding-plane slip. S1 axial plane cleavage is preferentially developed in folded shale layers. In fold limbs S1 is parallel to S0 and produces a composite S0/S1 foliation. S0/S1 is defined by compositional layering that reflects the original compositional variation. Fold orientations are variable as a result of reorientation by later folding. Reoriented $F1_{EZ}$ fold axes plunge steeply towards the NW to NE with generally steep to moderately dipping axial planes that dip to the SE, NE, and NW.

$F2_{EZ}$ folds reorient S0/S1 and are inclined to recumbent with a well developed S2 axial plane cleavage. $F2_{EZ}$ fold axes plunge moderately to shallowly to the N to NE with generally NE- to NW-striking axial planes that dip moderately to the E to SE and NE. S2 axial plane cleavage is the most prominent cleavage developed in the Chignecto domain. A stretching lineation is associated with $F2_{EZ}$ and is reoriented by later folding.

$F3_{EZ}$ structures are the most prominent folds in the external zone. They reorient $F1_{EZ}$ and $F2_{EZ}$ folds and their associated cleavages (S0/S1 and S2). $F3_{EZ}$ folds are upright with fold axes that plunge shallowly to moderately to the N to NE and SW. They have a distinct asymmetry wherein the long limbs strike N to NE and dip moderately to steeply to the E to SE. The short limbs, which generally have a shallower dip, strike NE and dip to the NW. The long limbs have been highly extended during folding. A typical fold axis plunge is 29º towards trend 028º. Axial planes strike N to NE and dip steeply to the E, SE, and NW. S3 axial plane cleavage is not well developed. Axial plane traces intersect the fault zone at a low angle.

The youngest folds ($F4_{EZ}$) are broad warps with axial plane traces that intersect the fault trace at a high angle. $F4_{EZ}$ folds are readily identified on aerial photographs and have NNE- and SSW-plunging fold axes and axial planes that strike NNE and dip steeply to the ESE and WNW. The long limbs of these folds strike NE and the short limbs strike W to NW.

Stretching lineations are associated with $F2_{EZ}$ folds and plunge shallowly NE to E and SW to WNW. The principal lineation orientation is a plunge of 09º towards 075 º.

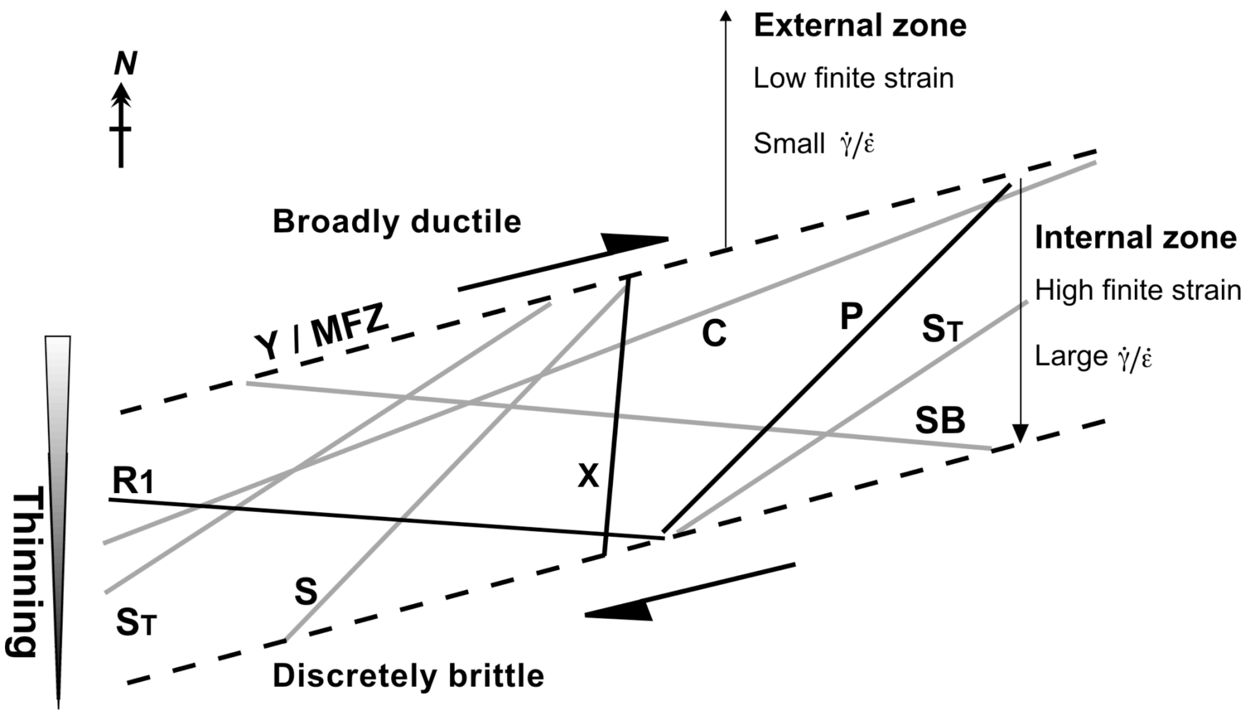
Internal Zone

The high-strain internal zone (exposed along Greville Bay) accommodates primarily transcurrent motion and can be further subdivided into three domains that reflect variations in lithology, finite strain, and style of deformation: (1) the marginal domain, (2) the core domain, and (3) the shear-band domain (Fig. 5-1). Figure 5-4 schematically shows the orientations of various planar structures that are developed in the internal zone.

Marginal Domain

The boundary between the external zone and the marginal domain of the internal zone is well defined. The amount of deformation accommodated in the marginal domain is significantly greater and more localized than in the external zone. The deformed rocks of the Greville River Formation have a strong compositional layering formed by layer transposition that obliterated the original sedimentary features. Black phyllite is the principal deformed unit. Multiple fold generations record a progressive and repeated rotation of fabrics into parallelism with the MFZ trace. The degree of parallelism increases with proximity to the contact with the core domain.

Fig. 5-4*:* *Major surfaces in the internal zone. Brittle structures shown by heavy lines; ductile structures shown in light lines.*



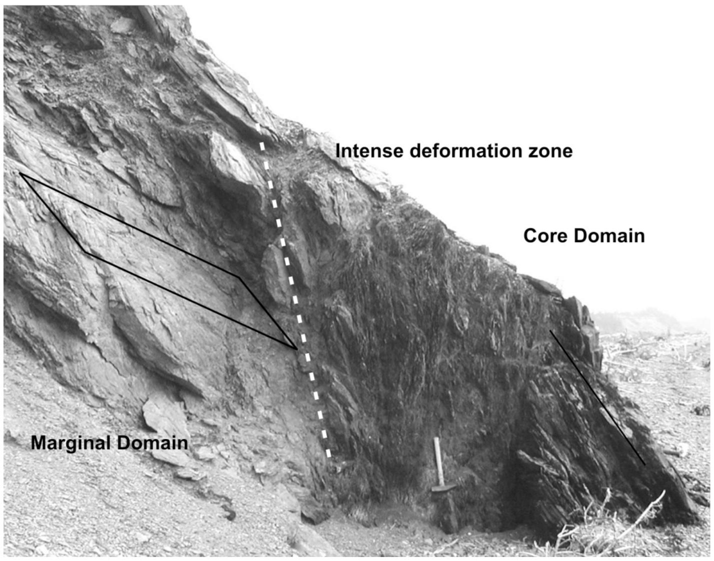
Scale independent schematic representation of the major surfaces and foliations from the internal zone. Brittle structures; Minas Fault Zone (MFZ), Riedel fault (R_1), X-fault, Y-plane, P-orientation. Ductile structures; S-foliation, C-plane, S_T- main compositional layering, shear-bands (SB).

Strain partitioning and localization is a dominant attribute of this high-strain zone. The kinematics are constant from low to high strain. The style of deformation varies from low to high strain. The ratio of the simple shear component (γ) to pure shear component (ε) also varies across the zone, simple shear is concentrated within the core domain of the internal zone. Finite strain is increasingly localized into narrower domains (thinning) which accommodate discrete dextral strike-slip displacements.

Core Domain

The primary differences between the core and marginal domains are a discontinuous increase in the level of deformation (Fig. 5-5), the strength and ubiquity of stretching lineations and the style of structures (Fig. 5-6). The core domain accommodates intense deformation in a relatively narrow zone (~ 100 m). Folding is intense and ductile fabrics of different generations tend toward a high degree of parallelism. Multiple veining is locally so intense that the bulk composition of fault rocks can be significantly altered. The style of subsequent deformation is largely controlled by the presence of these veins. There is abundant evidence for dextral strike-slip faulting that overprints earlier ductile fabrics. Fault-bounded slices of basement and Upper Carboniferous units are internally deformed and thinned.

Fig. 5-5: *Intense deformation zone (steep cleavage) in Greville River Formation, marks contact between marginal and core domains. Photo facing ENE along strike of the internal zone. Greville Bay.*



Shear-Band Domain

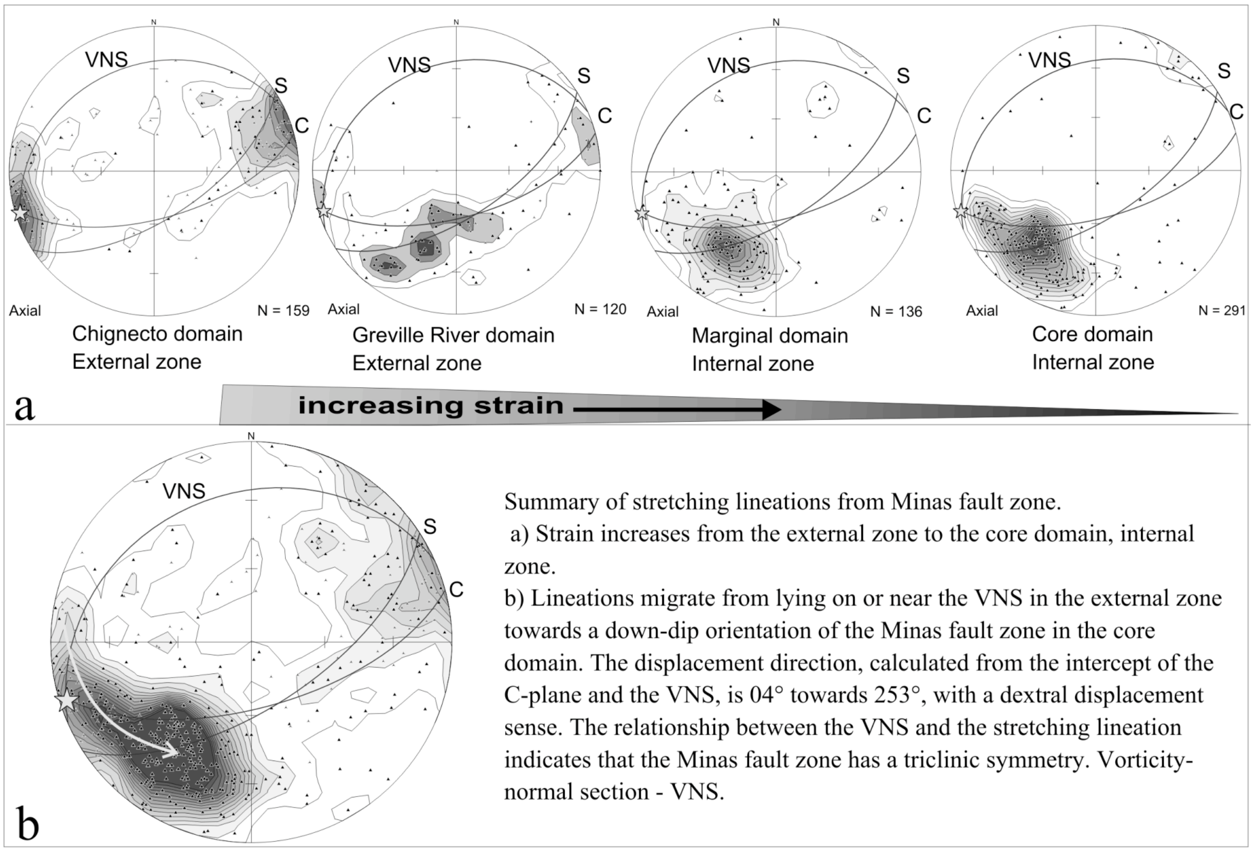
The shear-band domain, embedded in the core domain, is discontinuous along strike. It is distinguished by intensive development of S/C foliations and extensional shear-bands in black phyllite. This domain is interpreted to reflect the highest strain achieved within the fault zone, as has been noted elsewhere by Passchier (1984).

On a gross scale C-planes exhibit shear displacement of the compositional layering. C-planes are defined by pressure solution cleavage, gouge zones, ultramylonite and fine-grained micas. The largest C-planes form surfaces that bound zones of particularly intense shear-band formation.

S-foliations are defined by; dismembered quartz veins, axial planes of crenulated micas, and rotated compositional layering. They are oriented at an oblique angle to the shear zone boundary. The overlap of the S-foliation and the transposed foliation orientations reflects the fact that S-foliations form at a wide range of scales during heterogeneous deformation. With increasing deformation, S-surfaces rotate into parallelism with C-surfaces to produce a composite S/C fabric.

Shear-bands (Platt & Vissers 1980) are extensional features with discrete, penetrative surfaces that form at angles less than 35º to the shear zone margin and curve into the C-orientation. Displacement on the shear-bands is consistently dextral with corresponding offsets of the compositional layering, S-foliation, and C-planes. Individual shear-bands range in size from 1-15 cm. During progressive deformation shear-bands rotate into near parallelism with the shear plane and are subsequently overprinted by younger sets of shear-bands. The calculated shear direction throughout the internal zone is dextral with a small WSW-plunge; this relationship holds for both averaged regional orientation data (marginal and core domains) and more localized data (shear-band domain).

Fig 5-6: *Summary of stretching lineations from Minas Fault zone*



Summary of stretching lineations from Minas fault zone.
a) Strain increases from the external zone to the core domain, internal zone.
b) Lineations migrate from lying on or near the VNS in the external zone towards a down-dip orientation of the Minas fault zone in the core domain. The displacement direction, calculated from the intercept of the C-plane and the VNS, is 04° towards 253°, with a dextral displacement sense. The relationship between the VNS and the stretching lineation indicates that the Minas fault zone has a triclinic symmetry. Vorticity-normal section - VNS.

Interpretation

The preserved record of MFZ evolution (Fig. 5-2) illustrates both distinct lateral and vertical deformation distributions. The vertical variation in the distribution and accommodation of deformation is characterized by basin formation in the strata, which overlies the internal zone that accommodated large amounts of strain by transcurrent displacement. During half-graben formation, stratigraphic units of different ages were juxtaposed and incorporated into the fault zone.

The lateral distribution of deformation levels is what defines the different domains along the Cobequid fault zone. Deformation partitioning is a characteristic of transpression zones (Jones & Tanner 1995, Lin et al. 1998). Initially, deformation is broadly distributed over a wide area; with increasing strain, deformation is partitioned into increasingly narrow domains.

Localization ultimately results in the formation of the shear-band domain, which accommodates the highest levels of strain (Passchier 1984), and is geometrically associated with brittle and semi-brittle fault zones. With this increase in strain from edge to centre of the MFZ, there is a related progressive localization of deformation, culminating in the discrete shear-band domain. Domain widths decrease through the sequence (external zone – marginal domain – core domain – shear-band domain). The narrowing of definable domains (strain localization) and the concomitant increase of strain within the domains can be explained in terms of variations in strain rate throughout the deformation zone. In other words, because the accumulation of strain is distinctly heterogeneous, at least some component of finite strain variation is a function of strain-rate variation. The development of faster strain rates in order to accommodate imposed boundary condition displacements would have affected both work hardening rates and the degree of strain localization. Such strain-rate localization contributes to the transition from ductile to brittle structures. The similarities of the shear-band domain fabrics and the Riedel fault pattern orientations supply evidence for the latter type of transition (Fig. 5-4).

The quasi-parallelism between *C/S/shear-band* ductile fabrics and *Y/P/R1* brittle fabrics supports the contention that they fulfill the same mechanical and kinematical purpose (White et al. 1980, Mawer & White 1987). Although the faults could be argued to simply reactivate pre-existing ductile fabrics, the fact that they overprint all rock types would favour their origin as kinematic equivalents of the shear-band domain structures under conditions of extreme work hardening. Additionally, those cases where C/S fabrics overprint brittle deformation patterns demonstrate that there can be concomitant operation of large-scale brittle faulting and quasi-ductile deformation.

Ductile to brittle transitions are commonly the result of crustal exhumation or some other cause of crustal cooling. However, the MFZ fault rocks do not show any explicit mineralogical evidence consistent with extensive uplift through the ductile-brittle transition. The kinematic analysis has demonstrated that the MFZ is a thinning deformation zone in which strain is accommodated in progressively narrower volumes of rock. The basin in which lower Carboniferous Horton Group rocks were deposited has been collapsed into a much narrower zone between the Meguma and Avalon terranes. The asymmetry of deformation across the basin is striking, seen as the localized strain along the MFZ (northern Minas shore) and the well-preserved traspressional folding evident along the southern Minas shore. The narrowing of the deformation zone can be anticipated to have given extensive hardening consistent with progressive localization. In effect, the MFZ can be described as an initially diffuse (wide) zone of transpression comprising significant folding and thrusting typified by the external zone. With continued displacement, the MFZ evolved towards more localized zones of concentrated deformation related to shear zone thinning and the accumulation of finite strain. Such a progressive evolution makes the definition of a discrete "shear zone boundary" difficult. Specifically, it is not possible to demarcate "deformed shear zone" and "undeformed host rocks" and instead, the MFZ is described in terms of discontinuous transitions in finite strain and deformation style within a grossly common movement picture.

GREVILLE BAY

Access

If travelling from Stellarton, continue west on highway 104, following signs to New Brunswick. If travelling from Truro or Halifax, take highway 102 north and exit left when road divides at intersection with highway 104, following signs to New Brunswick.

0.0	*On-ramp from highway 102 joins highway 104 westbound.*
12.4	*Take exit 12 towards Parrsboro*
13.5	*Turn right at intersection with highway 4, towards Parrsboro*
15.7	*Turn left on highway 2 towards Economy and Parrsboro*
38.3	*Bass River (bridge in centre)*
59.3	*Entrance to Five Islands provincial park*
59.3	*Return to highway 2; turn left toward Parrsboro*
85.5	*Intersection in centre of Parrsboro - turn right, following highway 2*
87.7	*Cross Roads Garage: Turn left on highway 209 towards Advocate. Travel west on highway 209 which follows the scarp of the Cobequid fault. As one turns on to highway 209, there is a rifle range on the right behind which is a large fault block of the Jeffers Formation. At Greville Bay we will see other Jeffers Group fault blocks which have been incorporated into the fault zone. On the 25 minute journey to Greville Bay, there are isolated outcrops of deformed Greville River Formation rocks on the right side of the road. The foliation typically dips moderately to steeply to the southwest*
106.1	*Port Greville - Age of Sail Museum*
111.7	*Turn left on small dirt lane*
112.2	*Park at blue and white house and walk to shore along path. (Please inform occupant of your presence if not previously arranged. There have never been access problems and we wish to maintain this relationship.) We will park the vehicles and walk 300 m to the beach. Distances between outcrops are measured from the parking area.*

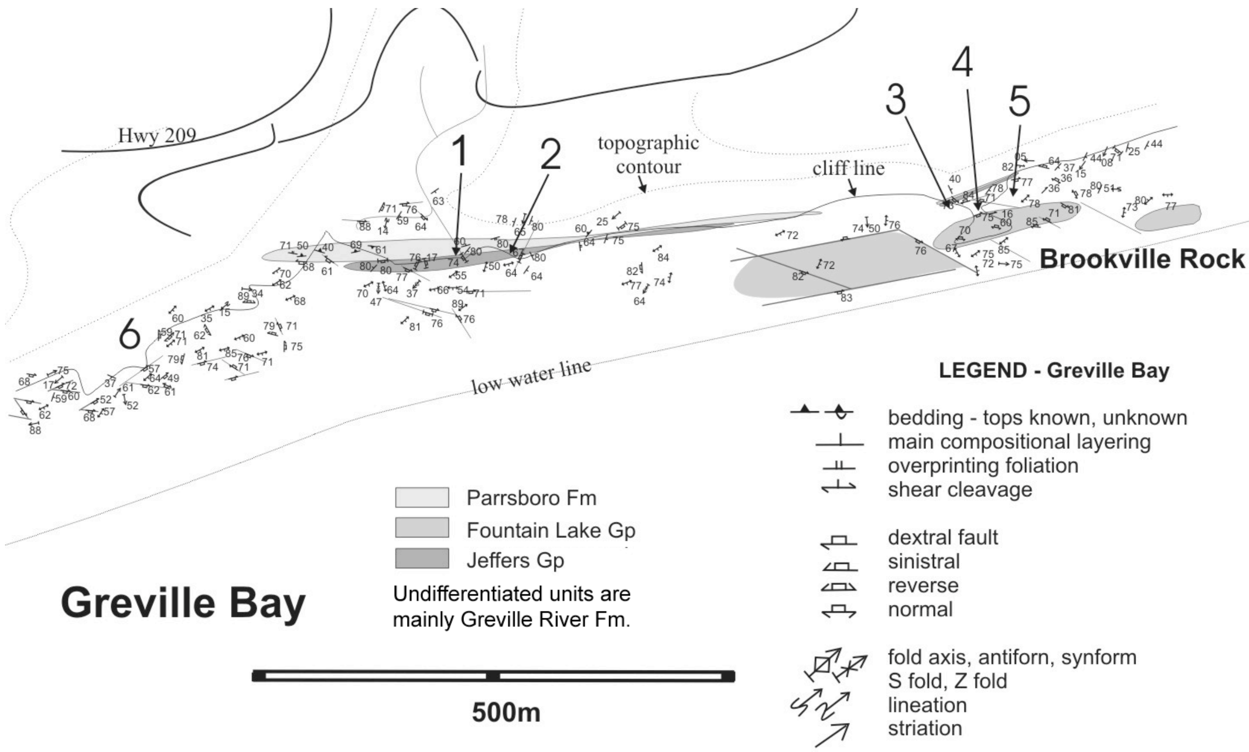
Upon reaching the end of the path to the beach, turn left (east) and cross a small stream toward the first exposure visible at the top of beach. Looking across the bay (if it is a clear day!) you can see the North Mountain Basalt headlands of Cape Blomidon and Cape Split on the Minas southern shore. Fig. 5-7 is simplified map of Brookville area with excursion locations marked and Fig. 5-8 shows views respectively east and west along the beach with the described outcrops marked.

GR-1 (300 m) Fault Slices I

At this first stop, the Upper Carboniferous Parrsboro Formation (Cumberland Group) and Neoproterozoic Jeffers Group are in fault contact, separated by a gouge zone. This exposure lies within the marginal domain of the External Zone which extends along strike with the cliff face. Bedding and cleavage of the Parrsboro Formation are oblique to the trace of the fault zone Cleavage development in the Parrsboro Formation is likely to be related to folding as seen approximately 75 m east of GR-2. The beds at GR-1 dip steeply to the southeast and are crosscut by steep fault planes with shallow striations plunging to the east consistent with dextral transcurrent displacement. Striations with a similar orientation exist within the Jeffers Group fault slice seen at GR-4.

The lithologically distinct fault slices provide important constraints on timing of deformation. Incorporation of the Jeffers Group fault slice into the fault zone may have occurred during N- to NW-directed thrusting dated as Mid Tournaisian to Mid Visean (Waldron et al. 1988), consistent with contractional deformation observed in the External Zone. The Jeffers Group is folded by $F3_{MD}$ and was therefore incorporated into the fault zone prior to this folding event. The strong compositional foliation in the Jeffers Group is parallel to the main trend of the fault zone and is axial planar to $F3_{MD}$.

Fig 5-7: *Map of shoreline west of Brookville Rock*



The incorporation of Parrsboro Formation into the fault zone provides some constraints on the timing of deformation. The Parrsboro Formation was deposited in a Late Carboniferous basin environment and subsequently incorporated into the fault zone. Therefore, the fault zone was active in a transpressive regime until the Late Carboniferous (Alleghanian event). The Parrsboro Formation is relatively undeformed compared to the older Horton Group, Fountain Lake Group, and Jeffers Group rocks, and has not been fully transposed into parallelism with the fault zone, suggesting that the transposition cycles were waning by this time, possibly due to a higher crustal level of deformation and/or rock type.

On the beach south of GR-1, the Jeffers Group is in fault contact with black phyllites of the Greville River Formation (Horton Group). As we proceed east towards GR-4 (Brookville Rock) we pass numerous steep faults that contain thin seams of gouge. Establishing a unique displacement sense on these faults is difficult, and it is likely that the faults in the cliff exposure are equivalents of strike-slip faults that are seen in beach exposures; alternatively they may be late extensional (Mesozoic) faults. The latter faults can be traced through the different fault slice lithologies and therefore are interpreted to postdate transposition and juxtaposition of these units. The compositional layering in the metavolcanic and metasedimentary units of the Jeffers Group is relatively well preserved compared to the Jeffers Group exposures at Brookville Rock. This foliation typically dips steeply northwest or southeast. Within the Jeffers Group block numerous veining episodes are recognized.

Fig. 5-8: *Location photos in Greville Bay*

GR-2 (400 m) Fault Slices II

Where the cliff cuts in to the north 100 m east of GR-1, the Parrsboro Formation sandstone is in fault contact with the Jeffers Group and the Greville River Formation. The near vertical fault planes which crosscut the Jeffers Group blocks have shallow striations. The main trace of the fault plane separating the units is not straight and has either been offset by these near vertical faults or curved as a result of anastomosing geometry or late folding.

Grey siltstones of the Greville River Formation can be seen high in the cliff. These strata exhibit a lower level of deformation than the same unit along the shore; i.e. they are farther from the focus of fault zone deformation, in the marginal domain as opposed to core domain of the Internal Zone. Isoclinal folds and steep, recognizable bedding planes characterize the deformation in the Greville River Formation. The bedding planes do not strike parallel to the fault zone, indicating a lower degree of transposition.

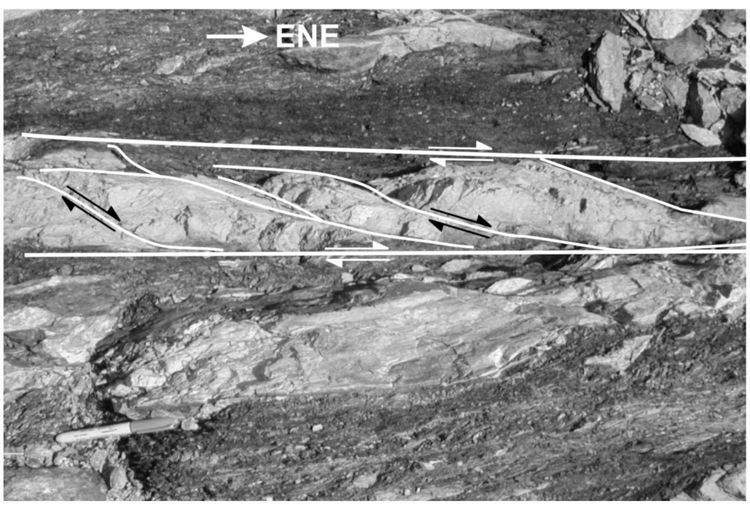
Folding within the Parrsboro Formation can be seen low in the cliff face approximately 75 m east of GR-1. Here the red and green siltstones show open recumbent folds with a later cleavage developed in the limb of the fold.

GR-2 to GR-3 (200 m – 800 m) Transposition and Fault Patterns

On the shore between **GR- 4 and GR-5 (subject to water level and seaweed concentrations)** the steeply dipping compositional layering (S_T) is deformed into open folds with moderate to steep plunges towards the SSW and steep axial planes dipping NNW. Earlier isoclinal fold closures can be seen within the limbs of these later folds. Discrete shears overprint these folds and offset their limbs.

The steeply dipping compositional foliation trends northeast, making angles of 30-45° to the main fault zone trace. This foliation is overprinted by faults that trend NW-SE and offset the layering in a dextral sense (Fig. 5-9). Within these 5-50 cm fault zones is an array of smaller oblique faults that also show dextral offsets.

Fig. 5-9: *Dextral Riedel shear offsets of competent sandstone layers in Greville River Formation*



The array of brittle faults reflects Riedel fracture patterns at different scales. At the fault-zone scale, the main fault zone boundary (Y) has oblique shears (R1) on the order of 100 m long that are synthetic with the dextral displacement sense on Y. Blocks between the large R1 faults are back-rotated by movement on R1, producing the 30-45° obliquity of foliations into the P-orientation. This geometric package of faults overprints the folding, and therefore records a late stage in the dextral displacement regime. Similar Riedel patterns are observed within the fault zone along individual 1st-order R1 faults. The partitioning of deformation within these scale-independent fault packages is demonstrated by the reverse sense of rotation of the P-oriented foliation resulting from displacement along the R1 shears. The preservation of these late-stage dextral shear indicators is dependent on cessation of displacements along the 1st-order faults.

The largest displacements in the earthquake cycle would have been accommodated along the Y-planes (parallel to the fault zone); the R1 and P orientations were obliterated in a process of resetting the fault zone to accommodate subsequent displacements.

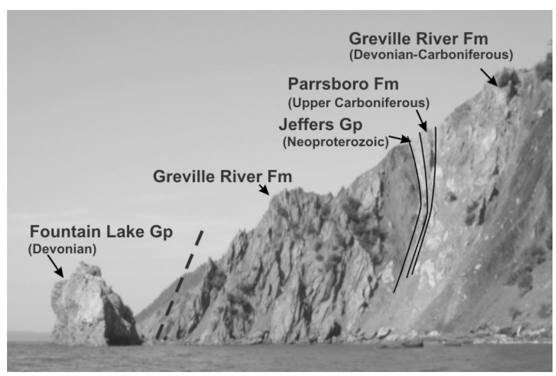
GR-3 (800 m) – Cliff cross-section on west side of Brookville Rock

In the cliff and promontory leading out to Brookville Rock, the transition from marginal to core domains of the Internal Zone is observed. Characteristic of the marginal domain, Greville River Formation strata in the cliff section are oblique to the main fault zone trace, and dip less steeply to the ESE. Shallow easterly-plunging striations can be seen high up on the fault surfaces of the green Jeffers Group and the yellow-weathering Parrsboro Formation slices, consistent with strike-slip displacement.

Moving toward the core domain of the fault zone, lithologically distinct fault slices stand out in the cliff sections and are deformed into parallelism with the fault zone (Fig 5-10). Note that the green Jeffers Formation slice is considerably thinner and the compositional layering less well preserved than the same unit that crops out in the section between GR- 1 and GR-3. The latter indicate a higher level of strain resulting from the transposition of the Jeffers Group slice into parallelism with the fault zone.

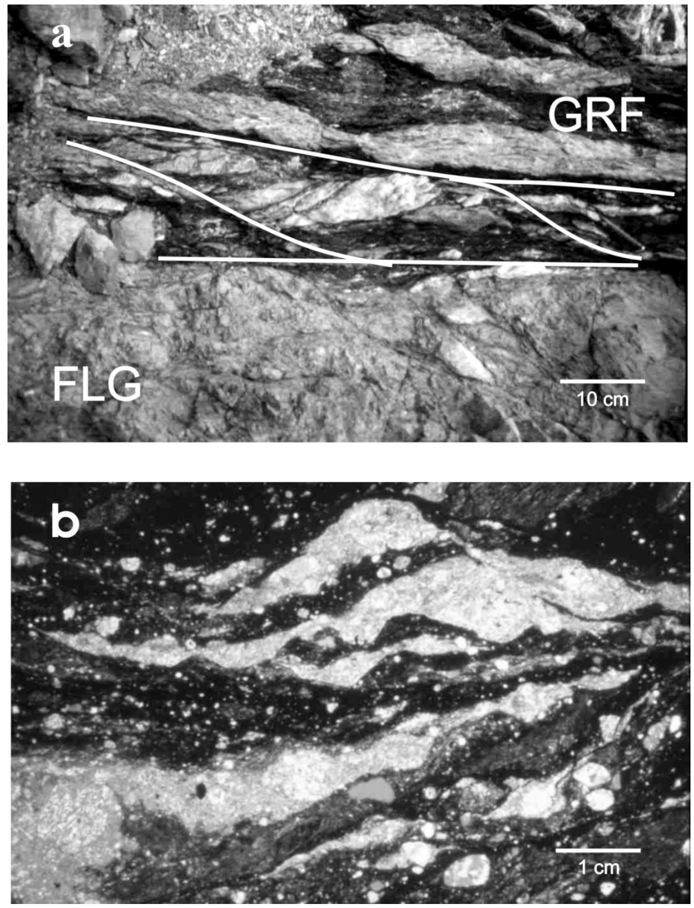
An intensification of deformation in the Greville River Formation, Parrsboro Formation and Jeffers Group is evident as the resistant Fountain Lake Group rhyolite (Brookville Rock) is approached. This reflects the role that local boundary conditions had on the partitioning of deformation and emphasizes the heterogeneous nature of shear zones. At this locality the phyllites and Jeffers Group have been highly deformed relative to the rhyolite. The deformation has been partitioned into the rock units which can more readily accommodate higher levels of strain due to their rheological properties.

Fig. 5-10: *GR-4 (1000 m) Brookville Rock viewed from east side looking west showing lithostratigraphic zonation in fault zone.*



As we walk between the grey phyllites of the Greville River Formation and the Fountain Lake Group rhyolite block which forms Brookville Rock you will notice an exposure of green and black cataclasites, gouges and phyllonites which preserve excellent semi-brittle deformation features (Fig. 5-11). Riedel fracture patterns can be seen on various scales and indicate dextral shearing. Lineations can be seen on the surfaces of fault blocks within this ~2 m deformation zone. These striations have shallow plunges to the east and correspond to those seen on the Jeffers Group and Parrsboro Formation exposures in the cliff face (GR-3). The Riedel patterns are transitional with S-C and shear band fabrics observed in slightly more ductile zone.

The more competent rhyolite block is massive to locally foliated and shows considerably less evidence of high strain than the more incompetent phyllites of the Greville River Formation. High angle faulting and breccia zones characterize the brittle deformation within the rhyolite block.



5-11: *Systematic shear patterns and foliation in localized semi-brittle deformation zones, Brookville Rock. (a) Outcrop structures GRF- Greville River Formation; FLG - Fountain Lake Group. (b) Thin Section, Greville River Formation. Calcite-quartz fragments in phyllonitic matrix.*

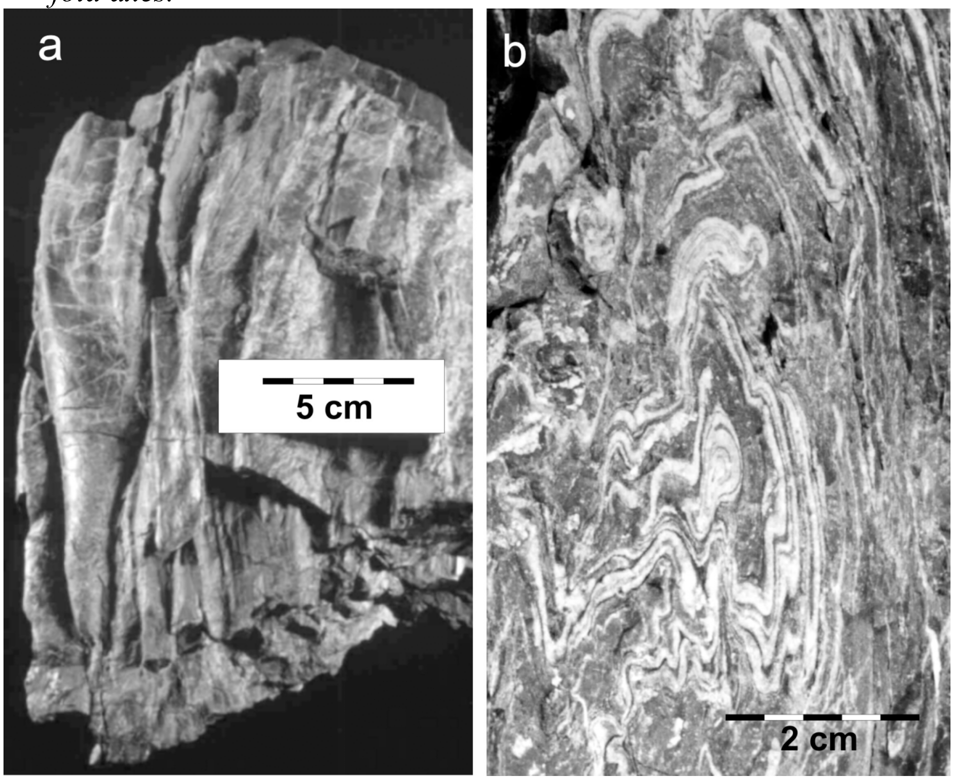
The outcrops of rhyolite along the beach to the west of the large rhyolite block are offset by faults in a Riedel shear orientation (R1) relative to the main fault zone (Y). The R1-shears exhibit dextral offsets and dip steeply to the south east. These Riedel shears can also been identified in the Greville River Formation and because they have been preserved (i.e. are not transposed into parallelism with the fault zone) may be a result of the latest 'locking up' phase of movement on the fault zone.

GR-5 (1020 m) - East Side of Brookville Rock

Proceeding past Brookville Rock to the eastern side of the promontary , it is possible to examine more closely the deformation in the Greville River Formation phyllonite. Sheath folds have been found in this horizon, defined by thin carbonate-rich layers and deformed syn-deformational veining (Fig. 5-12). The sheath folds have subhorizontal axes parallel to the main fault zone and have to date only been found along strike from this location in this distinctive finely laminated fault rock. Clearly the fault zone experienced high ductile strains during its tectonic evolution. Overprinting by semi-brittle and brittle deformation features is due to changing conditions during progressive transpression.

The shears within the Greville River Formation show dextral displacements and correspond to those to be seen at GR-6A. This is an interesting stop at which to consider the recognition of the boundary between ductile and brittle deformation.

Fig. 5-12: *Sheath folds with subhorizontal axes in fine-grained carbonate-quartz-mica schist, Greville River Formation. (a) Sheath fold surfaces. (b) Section perpendicular to fold axes.*



Looking to the east along the strike of the fault zone you can a see a distinct change in the types of fault rocks found and the structures produced. Later, large-scale folding of the fine-grained micaceous phyllites produces relatively open (non-isoclinal) folds which plunge moderately towards the south with axial planes dipping steeply to the east and west (at a high angle to the fault zone).

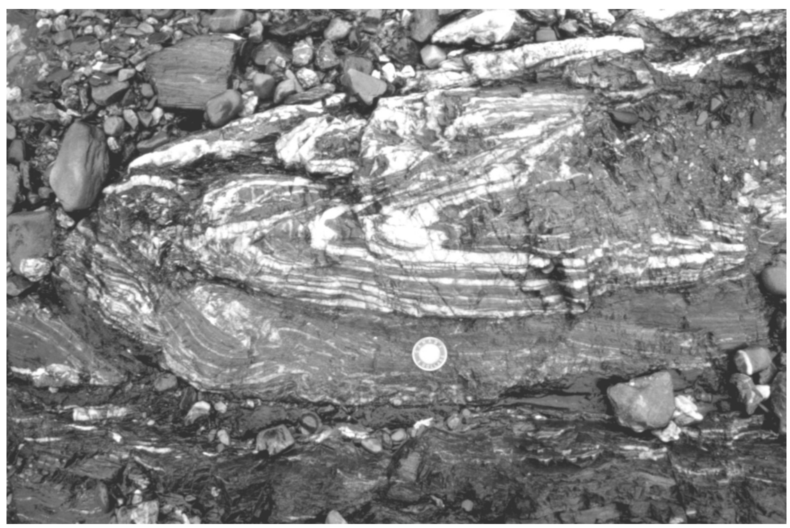
Return west along the beach, across the stream at bottom of beach access path, toward the outcrops on the western side of Brookville beach. Between the stream and the cliff outcrops, beach outcrops of Greville River Formation commonly show classic transposition and refolding structures. [Exposure of these low, subhorizontal outcrops varies dramatically from summer to summer as a function of winter storms, and whether erosion or deposition is dominant.]

GR-6 (2040 m; 300 m from GR-1) Core Domain Deformation

In following the cliff outcrops toward a small, but prominent sea stack (Fig. 5-8), there is an increase in finite strain that is consistent with a focussing of the deformation into narrower zones. This transect crosses through the core domain of the fault zone and ends in the shear-band zone at the sea stack. Apparent in these outcrops are the difficulties encountered in discriminating among distinct fault rocks derived from the same Greville River Formation protolith (MacInnes 2005). Distinct fault rock types can be mapped over tens of kilometres of fault zone length, and are central to establishing the structural geometry and kinematics. The exposures are described in three main blocks (Fig. 5-8).

Block A consists of alternating green and black phyllites. Within this block, intrafolial fold hinges are preserved, their limbs attenuated during transposition. At least three generations of progressive intrafolial folds (Fig. 5-13) have been identified in this green and black unit, indicative of the cyclical nature of the transposition which repeatedly rotates pre-existing structural elements into parallelism by isoclinal folding. The plunges of isolated isoclinal fold hinges vary depending on the intensity of transposition and number of cycles experienced. Looking west, along the strike of the fault zone (at a vertical surface), one can see examples of refolded isoclinal folds in the black and green phyllite. Quartz veins are aligned parallel to the axial plane of the fold and some appear to have been refolded. The transposed foliation has been crenulated. Transposition can be viewed at mesoscopic scale (by folding of the transposed foliation) and on a thin section scale (isoclinal folds). Progressive non-coaxial straining of the rocks to large magnitudes of strain produced a quasi-steady-state transposition foliation seen as the dominant compositional foliation. Transposition and the added effect of cyclical dynamic recrystallization make it impossible to unambiguously determine how many cycles of deformation the phyllite has undergone and therefore impossible to determine the strain magnitudes.

Fig. 5-13: *Refolded quartz veins in black phyllite, Greville River Formation.*

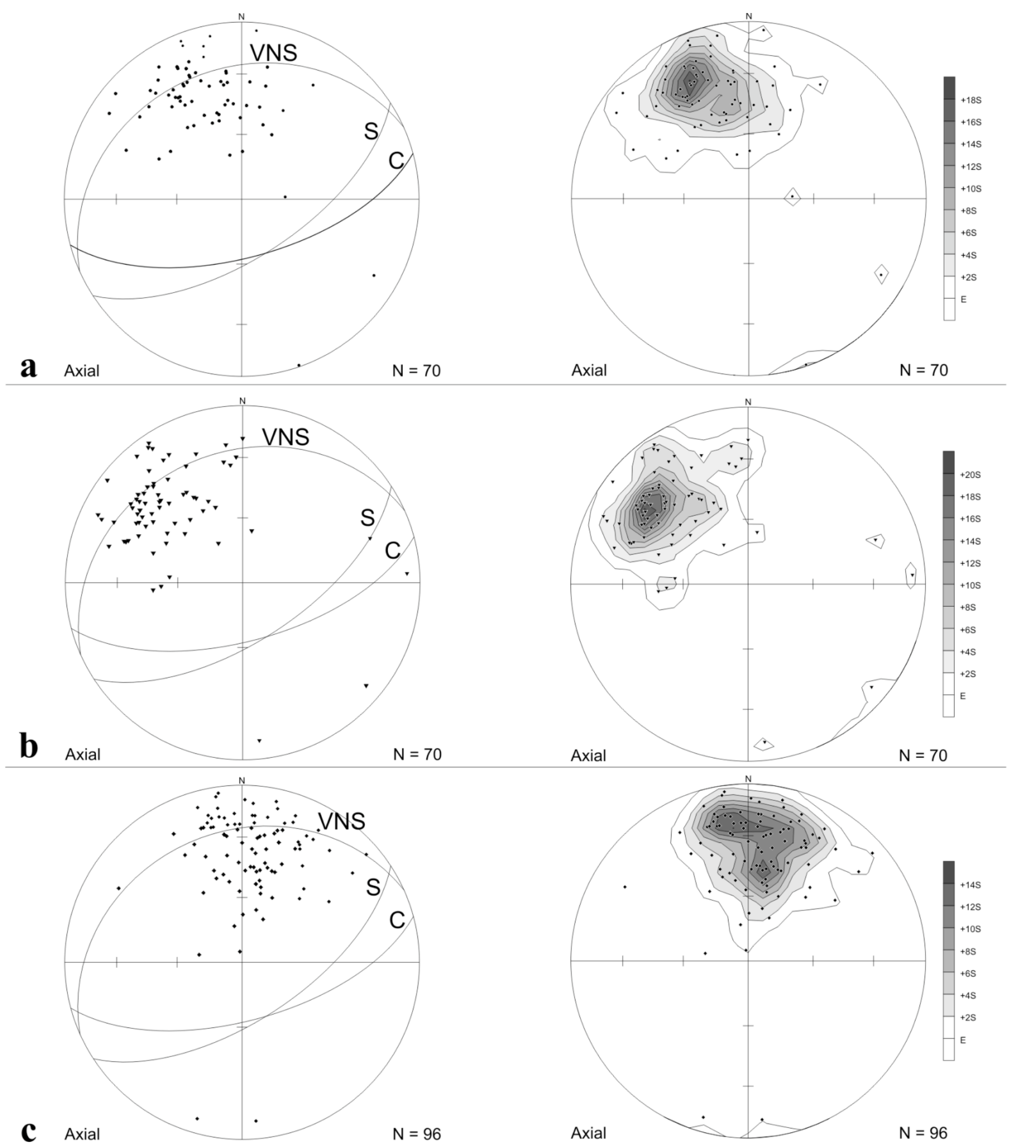


Block B lies immediately north of the sea stack (C). It is an L-S tectonite comprising 90% quartz and 10% phyllosilicates. The quartz displays intense intracrystalline deformation and has a strong foliation. S-folations occur as elongated/flattened quartz and the phyllosilicate seams form C-planes. The strong lineation plunges moderately to the southwest and is consistent with a high proportion of the plunges of intrafolial folds within the L-S tectonite and other fault rock types. Open, low-amplitude folds overprint the transposed foliation and intrafolial folds.

Chocolate tablet boudins occur at B. The stiff layer is extended in two directions within the layering, but predominantly towards 220°. Looking at the transposed layers in cross-section, isoclinally folded quartz layers can be seen to parallel the transposed foliation. Within this unit compositional variations can be seen within the boudins when viewed perpendicular to the transposed foliation and down the plunge of the boudins (to the SW).

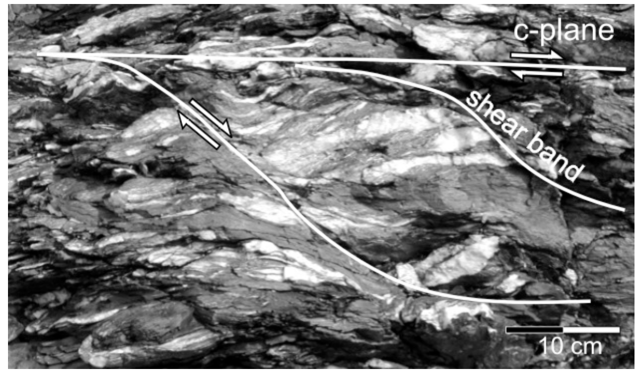
The sea stack (C) is typical of the shear-band domain that occurs discontinuously along strike within the core domain. The main compositional layering is formed by transposition of Greville River Formation concomitant with extensive and cyclic vein injection. On a gross scale C-planes exhibit shear displacement of the compositional layering. C-planes are defined by pressure solution cleavage, gouge zones, ultramylonite and fine-grained micas. C-planes in the core domain typically strike 069º and dip 57º SE (Fig. 5-14). The largest C-planes form surfaces that bound zones of particularly intense shear-band formation.

Fig. 5-14: *Equal area lower hemisphere projection of S-foliations, C-planes, and shear bands from the shear-band domain. VNS- vorticity-normal section, average C-plane, and S-foliation shown. (a) Poles to C-planes. (b) Poles to S-foliations. (c) Poles to shear-bands. Bar to right of strereonet defines contour density.*



S-foliations are defined by: dismembered quartz veins, axial planes of crenulated micas, and rotated compositional layering (Fig. 5-15). They are oriented at an oblique angle to the shear zone boundary and typically strike 044º and dip 57ºSE (Fig. 5-14). The overlap of the S-foliation and the transposed foliation orientations reflects the fact that S-foliations form at a wide range of scales during heterogeneous deformation. With increasing deformation, S-surfaces rotate into parallelism with C-surfaces to produce a composite S/C fabric.

Fig. 5-15: *Shear bands deforming transposed quartz veins in phyllonite in core domain of fault zone.*



Shear-bands (Platt & Vissers 1980) are extensional features with discrete, penetrative surfaces that form at angles less than 35º to the shear zone margin and curve into the C-orientation. Displacement on the shear-bands is consistently dextral with corresponding offsets of the compositional layering, S-foliation, and C-planes. Individual shear-bands range in size from 1-15 cm. The typical shear-band strikes 095º and dips 57º S (Fig. 5-14). During progressive deformation shear-bands rotate into near parallelism with the shear plane and are subsequently overprinted by younger sets of shear-bands. The calculated shear direction throughout the internal zone is dextral with a small WSW-plunge; this relationship holds for both averaged regional orientation data (marginal and core domains) and more localized data (shear-band domain).

With the increase in strain from the marginal to core domain, there is a localization of deformation and a decrease in domain width as strain is accommodated into progressively narrower volumes of rock, culminating in the discrete shear-band domain. Overprinting relationships demonstrate a general progression from ductile through semi-brittle to brittle. The quasi-parallelism of the shear-band domain fabrics (*C/S/shear-band ductile fabrics*) and the Riedel fault patterns *($Y/P/R_1$ brittle fabrics*) supports the contention that they fulfill the same mechanical and kinematic roles (Fig. 5-4). The fact that faults overprint all rock units suggests that they originated as kinematic equivalents of the shear-band domain structures under conditions of extreme work hardening.

Proceed back to beach access path and return to vehicles. Return to Parrsboro.

CLARKE HEAD

Access

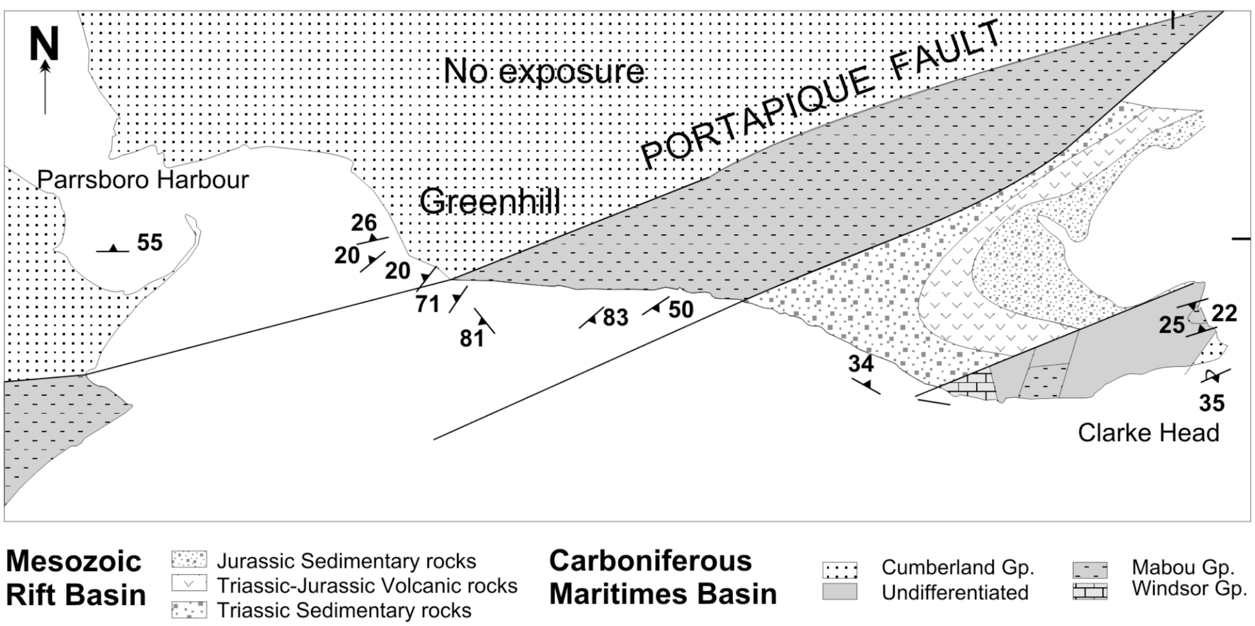
0 km *Fundy Geological Museum. From the museum parking lot, turn left and proceed the few metres to the junction with Two Islands Road. Turn right (east) on to Two Islands Road and proceed up the hill past the golf course to Glooscap Campground which is on the right (south) side of the road.*

4.5 km *Glooscap Campground. Park vehicles at campground entrance and proceed south through the campground to beach stairs and descend to the beach (500 m). Turn east (left) and proceed along the beach. Subsequent distances are measured from the beach stairs. We will walk east towards Clarke Head and will continue toward the headland for a distance of approximately 1300 m, then return by the same route to the vehicles.*

Overview

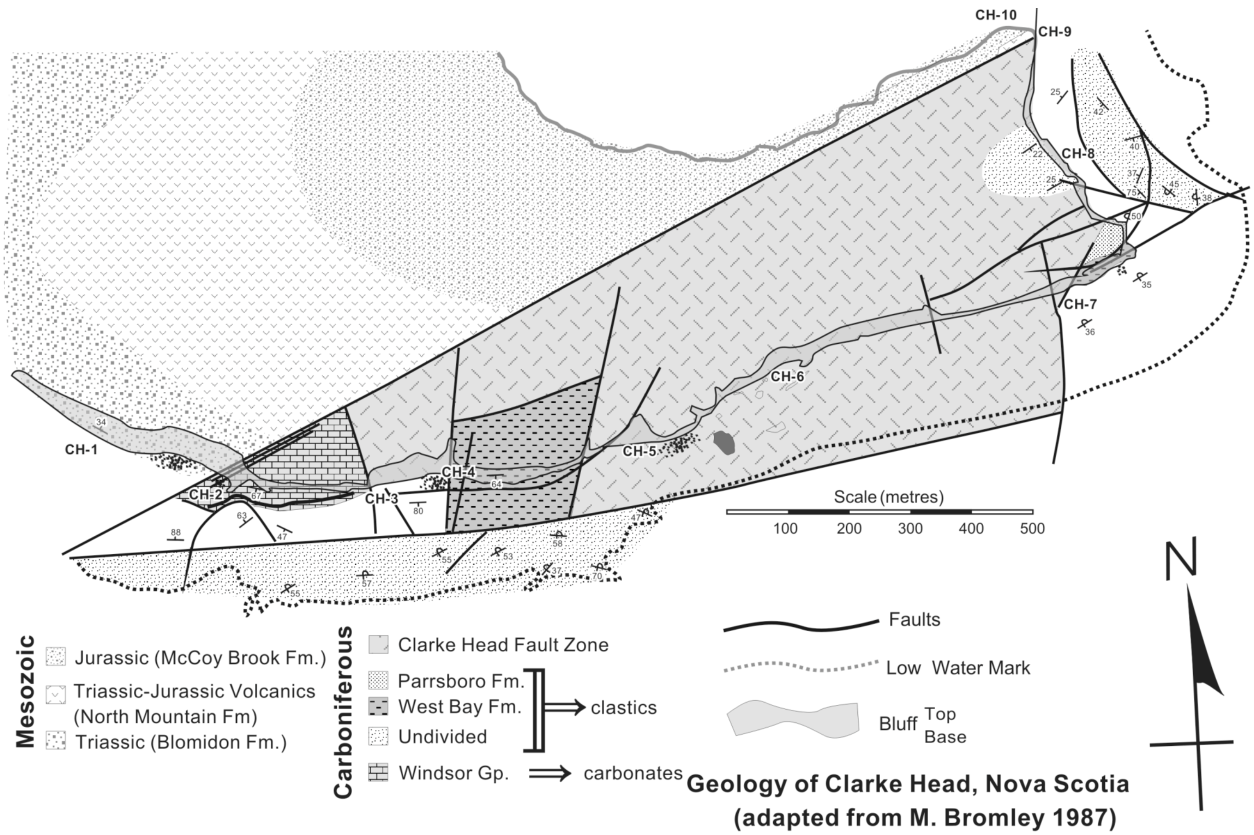
This excursion to Clarke Head (Figs 5-16, 5-17) examines ones of the most visited, yet enigmatic segments of the Minas fault system. This is the most outboard slice of the MFZ exposed along the northern Minas shore. Leaving Parrsboro we travel southeast across the Upper Carboniferous Cumberland Group (Parrsboro Formation) and the Middle Carboniferous Mabou Group (West Bay Formation).

5-16: *Geologic map of Parrsboro area.*



After parking the buses at Glooscap Park Campground, we will proceed to the stairs which lead down on to the beach. Basaltic headlands are prominent features along the Minas shore. To the south across the Minas Channel are the major Cape Blomidon and Cape Split basalt headlands comprising Triassic units of the Fundy Group, associated with Mesozoic rifting. From the head of the stairs there is a good view to the west of Partridge Island at the mouth of the Parrsboro River. This headland of North Mountain Formation basalt was separated from the mainland until a severe storm created a sand causeway joining them. The contact between the red sandstones of the Triassic Blomidon Formation and the overlying basalts can be seen.

***Fig. 5-17**: Map of Clarke Head.*



The excursion notes and map incorporate work from the unpublished thesis by Bromley (1987) with subsequent studies and observations (e.g. Gibbons et al. 1996).

CH-1 (200 m) Blomidon Formation

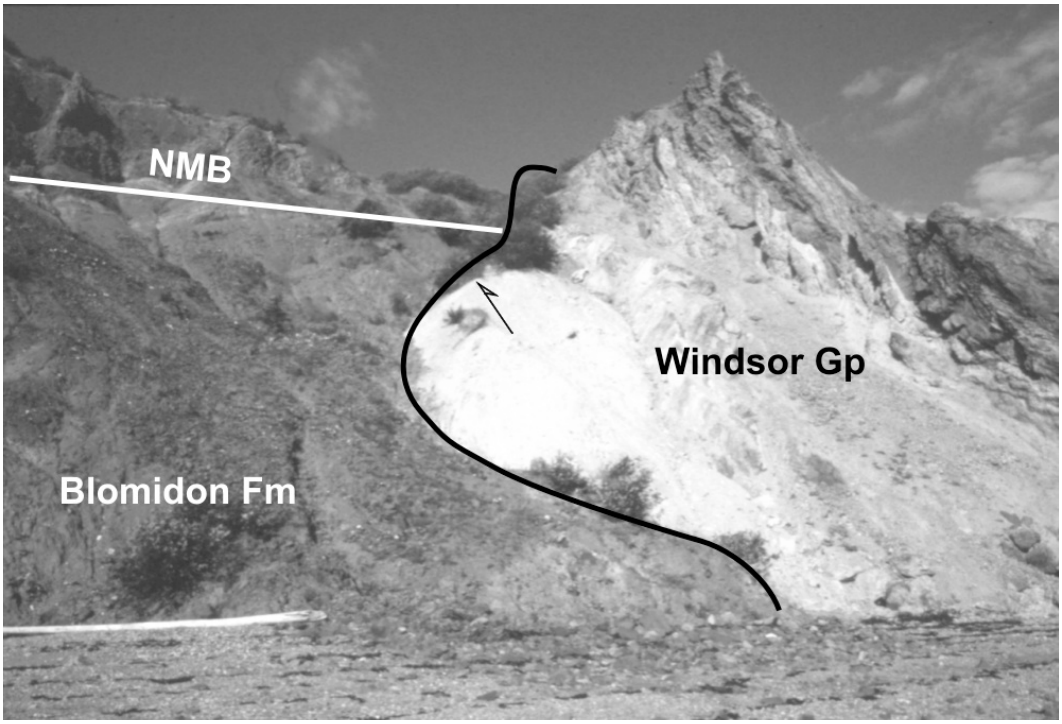
About 200 m east of the beach access, Triassic Blomidon Formation (Fundy Group) sandstones and siltstones, (Fig. 5-17) are exposed in the cliff, dipping shallowly to the northeast. Deposition of the Mesozoic units is related to formation of the basin during transtensional rifting of Pangea that formed the Fundy-Minas basins. The dominant deformation features at this location are micro- and meso-scale high-angle normal faults presumably associated with later post-Triassic extension.

CH-2 (400 m) Carboniferous-Triassic contact

At the first sharp bend in the cliff face, there are several striking stratigraphic and structural features. The Blomidon Formation sandstones are overlain at the top of the cliff by North Mountain Formation basalts that reflect the onset of rift volcanism (Fig. 5-18). Volcanic rocks are most extensively exposed on the south and east sides of Minas Basin and the Bay of

Fundy respectively, consistent with asymmetric rifting. The sandstone-basalt contact exhibits distinct colour variations as a result of contact metamorphism of the sediments.

Fig. 5-18 *Fault contact between the Windsor Group carbonates and the Triassic Blomidon Formation and overlying Jurassic North Mountain basalt (NMB). Photo facing east.*



These rift-sequence rocks are in steep fault contact with deformed Carboniferous (Namurian) upper Windsor Group limestone (Fig. 5-18). Fault gouge produces a dramatic colour contrast at the contact zone between the Triassic and the Carboniferous. The dominant mode of deformation within this relatively homogenous body of limestone is through fracturing and veining including tension gash arrays. Within the small overhang on the east side of this head land the limestone is folded and kinked. Gigantoproductid brachiopods are reported from the west end of the limestone outcrop. There is dispute as to whether the limestone unit is upright.

CH-3 (650 m) Windsor Group Brecciation

Rounding the headland and continuing east, the Windsor Group limestone terminates in fault contact with a zone of breccia about 100 metres wide. This zone contains blocks of shale, limestone, dolostone, serpentinite and gneiss supported by a light grey-green pulverised matrix. Along the base of the bluff is a distinctive band of dirty-orange limestone breccia about 2 metres wide which parallels the bedding in the cliff and is composed of limestone fragments in a limonitic limey matrix. The latter breccia appears to represent a bedding-parallel fault separating the Windsor Group from clastic Carboniferous strata that are observed in the intertidal zone.

CH-4 (750 m)

The eastern contact of the breccia zone is faulted against steeply dipping Carboniferous West Bay Formation strata that extend for approximately 225 metres. The beds are medium-grey, evenly-bedded, pervasively fractured fine-grained sandstones, shales and siltstones. Upright bedding surfaces form the face of the bluff and terminate in fault contact with the main mass of the megabreccia which is continuously exposed along the southern bluff for the next 600 metres eastward.

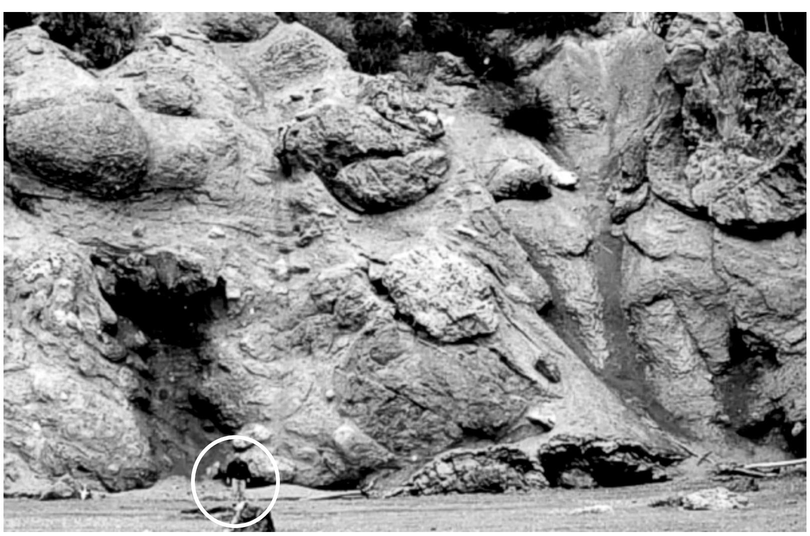
CH-5 (800 m) Megabreccia Zone

Upon entering the zone of megabreccia, a boulder field which has a high proportion of metamorphic rocks (serpentinite, marble, granulite gneiss) extends for 100 metres. The source of these boulders is the cliff immediately above, within the megabreccia. Standing in the intertidal zone, at the east end of the boulder field, is a large 25 m high stack of scapolite-rich metagabbro interpreted to be a faulted-in slice of Neoproterozoic Jeffers Formation. The granulite gneiss boulders (cpx-opx-plag-gnt) are the deepest-level rocks contained within the Minas Fault Zone. These exotic blocks are indicative of the complex and long-lived history of this terrane boundary (see CH-6).

CH-6 (1000 m) Clarke Head Megabreccia (Fig. 5-19)

Igneous, sedimentary and metamorphic basement rock fragments are found within the megabreccia, ranging in size from 10s of metres down to centimetres and are supported within a finer grained, quartzose, clayey (illite and chlorite) and gypsiferous matrix. Satin spar gypsum veins cross-cut blocks and matrix, while pyrite and large gypsum crystals are distributed throughout the matrix.

Fig. 5-19 *Clarke Head Megabreccia. Person for scale (circle)*



Time constraints on the tectonic history of the megabreccia (Gibbons et al. 1996) are strongly dependent on records within the granulite gneiss block. Granulite-grade mylonites within the megabreccia exhibit Devonian metamorphic zircon ages (369 Ma: U-Pb zircon) that are significantly younger than the Precambrian age of mainland granulites. This date is interpreted as the time of deep seated ductile shearing within the Minas fault system and is coeval with the initiation of widespread continental clastic sedimentation that produced the Horton Group. At this time the Minas Fault Zone may have acted regionally as the southern margin of a wide Maritimes Carboniferous basin.

Subsequent mid-crustal brittle deformation is recorded by hastingsitic amphibole veins (ca. 335 ± 3 Ma: $^{40}Ar/^{39}Ar$) which cut the mylonitic granulite fabric. These amphibole veins are the most Cl-rich amphiboles yet reported, comparable to hydrothermal systems such as the Salton Sea, California. This early Carboniferous age corresponds to uplift during post-Windsor Group marine regression (Yeo & Ruixiang 1987) and the establishment of fluvial conditions during Serpukhovian time as expressed by the Mabou Group.

A post-Mabou (Serpukhovian) age for the megabrecciation event is inferred from the abundance of Windsor limestones and evaporites and Mabou clastics within the megabreccia. Uplift and cannibalization of the earlier sedimentary infill produce the conglomeratic fluvial red beds of the Cumberland Group which overstep the steeply dipping and fault-brecciated Mabou Group. This Bashkirian to Moscovian event is 315-310 Ma and corresponds to the late stages of the Alleghanian orogeny.

Return by same route to vehicles

Although not part of this excursion, the following additional stops are included here in the case of extended future excursions.

CH-7 (1350 m)

Continuing east towards the headland, the megabreccia is seen to be in fault contact with the Late Carboniferous Parrsboro Formation (Cumberland Group) at the southeast corner of the Clarke Head peninsula. The Parrsboro Formation exists as a series of fault slices of continental clastic sediments with a total thickness of 400 m. The beds are overturned to the northwest and display mudcrack casts and sole markings. Passing around the headland, West Bay Formation (Mabou Group) is again encountered; the latter is overturned and underlies the Parrsboro formation.

CH-8 (1700 m)

Across the east end of the peninsula, there is a 300 m zone of megabreccia that is an along-strike of the megabreccia exposed at CH-6. Absence of outcrop on the surface precludes confirmation that the megabreccia zone is continuous, but it is nevertheless anticipated. Overlying the megabreccia is a sequence of red and grey-green, slightly calcareous siltstones and sandstones about 4 m in thickness. This unit appears to rest unconformably on the megabreccia, but it could also be a large slab included within it or even thrust upon it.

CH-9 (1900 m)

The megabreccia unit is in fault contact here with North Mountain Basalt which displays columnar structure in a subhorizontal attitude. The alteration zone between the columns contains large sheets of asbestiform anthophyllite.

CH-10 (1950 m)

Around the northeast corner of the peninsula, the North Mountain Basalt is in turn faulted against the McCoy Brook Formation (Jurassic), the youngest rocks observed in the excursion.

Looking across the bay towards the northeast, we see the Mesozoic red beds and basalts at Wasson's Bluff. This is an exciting palaeontological site famous for the discovery of extensive bone beds of "miniature" early dinosaurs. In the Fundy Geological Museum there is an exhibit on their recent work at Wasson's Bluff.

At the headland there appears to be a figure of a man leaning against the rocks. Local legend tells of this being the figure of Glooscap, the chief god of the Mi'kmaq, the indigenous inhabitants of Nova Scotia. You may have noticed the 6 m statue of Glooscap in Parrsboro town square. At Clarke Head, he is looking out towards the islands in the basin (appropriately named Two Islands and Five Islands) which he created by throwing sod at a beaver. Glooscap then threw jewels which explains why this area is so rich in minerals!

REFERENCES

Bell, W.A. 1927. Outline of Carboniferous stratigraphy and geologic history of the Maritime Provinces of Canada. Royal Society of Canada Transactions, Section IV, 75-108.

Boehner, R. C. 1991: Seismic interpretation, potential overthrust geology and mineral deposits in the Kennetcook Basin, Nova Scotia. Nova Scotia Department of Natural Resources, Mineral Resources Branch Report of Activities 1990, 37-47.

Boyle, R.W. 1957. Geology, Walton-Cheverie area, Nova Scotia. Geological Survey of Canada Bulletin 166.

Bradley, D.C. 1982. Subsidence in Late Paleozoic basins in the northern Appalachians. Tectonics 1, 107-123.

Crosby, D.G. 1962. The Wolfville Map Area, Nova Scotia (21H/1). Geological Survey of Canada, Memoir 325.

Culshaw, N., Liesa, M. 1997. Alleghenian reactivation of the Acadian fold belt, Meguma zone, southwest Nova Scotia. Canadian Journal of Earth Sciences 34, 833-847.

Donohoe, H., and Grantham, R.G. 1989. Geological Highway Map of Nova Scotia, second edition. Atlantic Geoscience Society, Halifax. AGS Special Publication No. 1.

Durling, P., Marillier, F. 1993. Tectonic setting of Middle Devonian to Lower Carboniferous rocks in the Magdalen Basin. Atlantic Geology 29, 199-217.

Eberz, G., Clarke, D. B., Chatterjee, A. K., and Giles, P. S., 1991, Chemical and isotopic composition of the lower crust beneath the Meguma lithotectonic zone, Nova Scotia; evidence from granulite facies xenoliths: Contributions to Mineralogy and Petrology, 109, 69-88.

Eisbacher, G. H., 1969, Displacement and stress field along part of the Cobequid Fault, Nova Scotia: Canadian Journal of Earth Sciences, 6, 1095-1104.

---, 1970, Deformation mechanisms of mylonitic rocks and fractured granites in the Cobequid Mountains, Nova Scotia, Canada: Geological Society of America Bulletin, 81, 2009-2020.

Ferguson, S.A. 1983. Geological Map of Hantsport Area, Nova Scotia (1:25,000). Nova Scotia Department of Natural Resources, Mineral Resources Branch, Map ME 1983-1.

Gibbons, W., Doig, R., Gordon, T., Murphy, B., Reynolds, P., White, J.C. 1996. Mylonite to megabreccia: Tracking fault events within a transcurrent terrane boundary in Nova Scotia, Canada. Geology 24, 411-414.

Gibling, M. R., 1995, Upper Paleozoic rocks, Nova Scotia, in Williams, H., eds., Geology of the Appalachian-Caledonian Orogen in Canada and Greenland F-1. Geological Society of America, The Geology of North America, 493-523.

Giles, S. and Boehner, R.C. 1982. Geological Map of the Shubenacadie and Musquodoboit (1:50,000). Nova Scotia Department of Natural Resources, Mineral Resources Branch, Map ME 1982-4.

Giles, S. and Lynch, G. 1994. Stratigraphic omission across the Ainslie Detachment in east-central Nova Scotia. In Current Research 1994-D, Geological Survey of Canada, 89-94.

Henderson, J. R., Wright, T. O., and Henderson, M. N., 1986, A history of cleavage and folding: An example from the Goldenville Formation, Nova Scotia: Geological Society of America Bulletin, 97, 1354-1366.

Hesse, R., and Reading, H.G. 1978. Subaqueous clastic fissure eruptions and other examples of sedimentary transposition in the lacustrine Horton Bluff Formation (Mississippian), Nova Scotia, Canada. In: Matter, A., Tucker, M (eds) Modern and Ancient Lake Sediments. International Association of Sedimentologists Special Publication 2, 241–257.

Johnson, R.J.E., Van der Voo, R. 1986. Paleomagnetism of the Late Precambrian Fourchu Group, Cape Breton Island, Nova Scotia. Canadian Journal of Earth Sciences 23, 1673-1685.

Johnston, S.K. 1999. Structural Geology of the Horton and Windsor Groups, Cheverie, Nova Scotia. B.Sc. Honours Geology/Geography Thesis, Saint Mary's University.

Jones, R.R. and Tanner, P.W.G. 1995. Strain partitioning in transpression zones. Journal of Structural Geology 17, 793-802.

Keen, C. E., Kay, W. A., Keppie, J. D., Marillier, F., Pe-Piper, G., and Waldron, J. W. F., 1991, Deep seismic reflection data from the Bay of Fundy and Gulf of Maine: tectonic implications for the northern Appalachians: Canadian Journal of Earth Sciences, 28, 1096-1111.

Keppie, J.D. 1982. The Minas Geofracture. In St.Julien, P., Beland, J. (Eds.), Major structural zones and faults of the Northern Appalachians. Geological Association of Canada Special Paper 24, 263-280.

---, 1993. Synthesis of Paleozoic deformation events in the Canadian Appalachians. Geologische Rundschau 82 , 381-431.

--- and Dallmeyer, R.D. 1995. Late Proterozoic collision, delamination, short-lived magmatism, and rapid denudation in the Meguma terrane (Nova Scotia, Canada): constraints from 40Ar/39Ar isotopic data. Canadian Journal of Earth Science 32, 644-659.

--- and Dostal, J. 1980. Paleozoic volcanic rocks of Nova Scotia. Virginia Polytechnic Institute, Department of Geological Sciences Memoir 2, 249-256

--- and Krogh, T.E. 1990. Detrital zircon ages from Late Precambrian conglomerate, Avalon Composite Terrane, Antigonish Highlands, Nova Scotia. Geological Society of America, Abstracts with Programs 22, 27-28.

--- and Krogh, T.E. 1999. U-Pb Geochronology of Devonian granites in the Meguma terrane of Nova Scotia, Canada: Evidence for Hotspot melting of a Neoproterozoic source. The Journal of Geology 107, 555-568.

---, Dostal, J., Murphy, J.B., Nance, R.D. 1996. Terrane transfer between eastern Laurentia and western Gondwana in the Early Paleozoic: constraints on global reconstructions. In Nance, R.D., Thompson, M.D. (Eds.), Avalonian and related peri-Gondwanan terranes of the circum North Atlantic. Geological Society of America. Special Paper 304, 369-380.

---, Dostal, J., Murphy, J.B., Cousens, B.L. 1997. Paleozoic within-plate volcanic rocks in Nova Scotia (Canada) reinterpreted: isotopic constraints on magmatic source and paleocontinental reconstructions. Geological Magazine 134, 425-447.

Kontak, D., Ansdell, K., and Douglas, A. 2000. Zn-Pb mineralization associated with a mafic dyke at Cheverie, Hants County, Nova Scotia: implications for Carboniferous metallogeny. Atlantic Geology 36, 7-26.

Krogh, T. E., and Keppie, J. D., 1990, Age of detrital zircon and titanite in the Meguma Group, southern Nova Scotia, Canada: Clues to the origin of the Meguma Terane: Tectonophysics 177, 307-323.

Lavoie, D. and Sangster, D.F. 1995. Origins and timing of basal Windsor carbonate breccias, Nova Scotia. In Geological Survey of Canada Current Research 1995-D., 79-88.

---, Sangster, D.F.; Savard, M.M. and Fallara, F. 1995. Multiple Breccia Events in the lower part of the Carboniferous Windsor Group, Nova Scotia. Atlantic Geology 31, 197-207.

Lin, S., Jiang, D., Williams, P.F. 1998. Transpression (or transtension) zones of triclinic symmetry: natural example and theoretical modelling. In Holdsworth, R.E., Strachn, R., Dewey, J.F. (Eds.), Continental transpressional and transtensional tectonics, Geological Society Special Publication 135, 41-57.

MacInnes, E. 2005. Structural and kinematic evolution of a segment of the Minas fault system, Nova Scotia, Canada. PhD thesis, University of New Brunswick, Fredericton, Canada.

--- and White, J.C. 2004. Geometric and kinematic analysis of a transpression terrane boundary: Minas fault system, Nova Scotia, Canada. In Alsop, G.I., Holdsworth, R.E., McCaffrey, K.J.W., Hand, M. (Eds). Flow Processes in Faults and Shear Zones. Geological Society of London, Special Publication 224, 201-214.

Marillier, F., and Verhoef, J., 1989, Crustal thickness under the Gulf of St. Lawrence, Northern Appalachians, from gravity and deep seismic data: Canadian Journal of Earth Sciences, 26, 1517-1532.

Martel, A. T. and Gibling, M. R., 1991, Wave-dominated shoreline facies and tectonically controlled cyclicity of the Lower Carboniferous Horton Bluff Formation, Nova Scotia, Canada, in Anadon, P., Cabrera, L., and Kelts, K., eds., Lacustrine Facies Analysis: Special Publication of the International Association of Sedimentologists 13, 223-243.

--- and Gibling, M. R., 1993, Clastic dykes of the Devono-Carboniferous Horton Bluff Formation, Nova Scotia: storm-related structures in shallow lakes: Sedimentary Geology 87, 103-119.

--- and Gibling, M. R., 1996. Stratigraphy and tectonic history of the Upper Devonian to Lower Carboniferous Horton Bluff Formation, Nova Scotia. Atlantic Geology 32, 13-38.

Mawer, C.K., White, J.C. 1987. Sense of displacement on the Cobequid-Chedabucto fault system, Nova Scotia. Canadian Journal of Earth Sciences 24, 217-223.

McCutcheon, S.R., Robinson, P.T. 1987. Geological constraints on the genesis of the Maritimes Basin, Atlantic Canada. In Beaumont, C., Tankard, A.J. (Eds). Sedimentary basins and basin-forming mechanisms. Canadian Society of Petroleum Geologists Memoir 12, 287-297.

Moore, R.G. 1986. Geological map of the Windsor area, Nova Scotia [NTS 21A16 north and 21H01 south] [1:25000]. Nova Scotia Department of Natural Resources, Mineral Resources Branch, Open File Map ME 1986-2.

---, 1993 (a). Geological Map of Cheverie-Lower Burlington Quadrangle, [NTS 21H01] [1:10 000]. Nova Scotia Department of Natural Resources, Mineral Resources Branch, Open File Map ME 1993-1.

---, 1993 (b). Geological Map of Cogmagun River-Goshen Quadrangle, [NTS 21H01] [1:10 000]. Nova Scotia Department of Natural Resources, Mineral Resources Branch, Open File Map ME 1993-2.

---, 1994. Geology of the Walton-Rainy Cove Brook map area [NTS 21H01-Z2 and Z4], Hants County, Nova Scotia. Nova Scotia Department of Natural

Resources, Mineral Resources Branch, Open File Report 1994-2.

---, 1996. Geology of the Cambridge Cove/Bramber and Red Head map areas, NTS 21H01-Z1, Z3 and 21H01-Y2. Nova Scotia Department of Natural Resources, Mineral Resources Branch, Open File Report ME 1996-2

--- and Ferguson, S.A. 1986. Geological map of the Windsor area, Nova Scotia. Nova Scotia Department of Natural Resources, Mineral Resources Branch, Map ME 1986-2.

Moore, R.G., Ferguson, S.A., Boehner, R.C. and Kennedy, C.M. 2000. Preliminary Geological Map of the Wolfville/Windsor Area, Hants and Kings Counties, Nova Scotia [21H/01 and parts of 21A/16C and D]. Nova Scotia Department of Natural Resources, Open File Map ME 2000-3.

Muecke, G.K., Elias, P., Reynolds, P.H. 1988. Hercynian/Alleghenian overprinting of an Acadian terrane: 40Ar/39Ar studies in the Meguma zone, Nova Scotia, Canada. Chemical Geology 73, 153-167.

Murphy, J.B., Hamilton, M.A. 2000. Orogenesis and basin development: U-Pb detrital zircon age constraints on evolution of the Late Proterozoic St. Marys Basin, central mainland Nova Scotia. The Journal of Geology 108, 53-71.

---, Keppie, J.D. 1998. Late Devonian palinspastic reconstruction of the Avalon-Meguma terrane boundary: implications for terrane accretion and basin development in the Appalachian orogen. Tectonophysics 284, 221-231.

---, Cameron, K., Dostal, J., Keppie, J.D., Hynes, A.J. 1985. Cambrian volcanism in Nova Scotia, Canada. Canadian Journal of Earth Sciences 22, 599-606.

---, Keppie, J.D., Hynes, A.J. 1991. Geology of the Antigonish Highlands, Nova Scotia. Geological Survey of Canada, Paper 89-10.

---, Rice, R.J., Stokes, T.R., Keppie, J.D. 1995. St. Marys basin, central mainland Nova Scotia: Late Paleozoic basin formation and deformation along the Avalon-Meguma terrane boundary, Canadian Appalachians. In Hibbard, J.P., van Staal, C.R. and Cawood, P.A. (Eds.), Current perspectives in the Appalachian-Caledonian Orogen. Geological Association of Canada, Special Paper 41, 409-420.

---, Keppie, J.D., Dostal, J., Waldron, J.W.F., Cude, M.-P., 1996. Geochemical and isotopic constraints on the accretion of the Avalonia in the Appalachian-Caledonide orogen: evidence from Early Silurian clastic sequences in Antigonish Highlands, Nova Scotia, Canada. Canadian Journal of Earth Sciences 33, 379-388.

---, Keppie, J.D., Davis, D., Krogh, T.E. 1997. Regional significance of new U-Pb age data for Neoproterozoic igneous units in the Avalonian rocks of northern mainland Nova Scotia, Canada. Geological Magazine 134, 113-120.

---, Keppie, J.D., Nance, R.D. 1999a. Fault reactivation within Avalonia: plate margin to continental interior deformation. Tectonophysics 305, 183-204.

---, Pe-Piper, G., Nance, R.D., Turner, D., Piper, D.J.W. 2000. Geology, eastern Cobequid Highlands, Nova Scotia. Geological Survey of Canada, Open- file 3703.

---, Fernandez-Suarez, J., Jeffries, T. E., and Strachan, R. A., 2004, U–Pb (LA–ICP-MS) dating of detrital zircons from Cambrian clastic rocks in Avalonia: erosion of a Neoproterozoic arc along the northern Gondwanan margin: Journal of the Geological Society, 161, 243-254.

Nance, R.D., Murphy, J.B. 1994. Contrasting basement isotopic signatures and the palinspastic restoration of peripheral orogens: Example from the Neoproterozoic Avalonian-Cadomian belt. Geology 22, 617-620.

O'Brien, S. J., Wardle, R. J., and King, A. F., 1983, The Avalon Zone: A Pan-African terrane in the Appalachian Orogen of Canada: Geological Journal 18,195-222.

O'Brien, S.J., O'Brien B.H., Dunning, G.R., Tucker, R.D. 1996. Late Proterozoic Avalonian and related peri-Gondwanan rocks of the Newfoundland Appalachians. In Nance, R.D., Thompson, M.D. (Eds.). Avalonian and related peri-Gondwanan terranes of the Circum North Atlantic. Geological Society of America Special Paper 304, 9-28.

Olsen, P. E., Schlische, R. W. 1990. Transtensional arm of the early Mesozoic Fundy rift basin: Penecontemporaneous faulting and sedimentation. Geology 18, 695-698.

Passchier, C.W. 1984. The generation of ductile and brittle shear-bands in a low-angle mylonite zone. Journal of Structural Geology 6, 273-281.

Piper, D. J. W. 1994. Late Devonian-Earliest Carboniferous basin formation and relationship to plutonism, Cobequid Highlands, Nova Scotia. In Current research, Geological Survey of Canada, 1994-D, 109-112.

--- and Pe-Piper. G. 2001. Tectonic deformation and magmatism along the southern flank of the Maritimes Basin: the northeastern Cobequid Highlands, Nova Scotia. Canadian Journal of Earth Sciences 38, 43-58.

---, Dessureau, G., Pe-Piper, G. 1999. Occurrence of early Carboniferous High-Zr rhyolites, Cobequid Highlands, Nova Scotia: temperature effect of a contemporaneous mafic magma. The Canadian Mineralogist 37, 619-634.

Platt, J.P., Vissers, R.L.M. 1980. Extensional structures in anisotropic rocks. Journal of Structural Geology 2, 397-410.

Plint, A. G., and Poll, H. W. van de, 1984, Structural and sedimentary history of the Quaco Head area, southern New Brunswick: Canadian Journal of Earth Sciences, 21, 753-761.

Roselli, C., 2003, Transpression and Overthrusting in the Minas Basin, Nova Scotia: M.Sc. thesis, University of Alberta, Edmonton AB Canada.

Ryan, R.J., Boehner, R.C., and Calder, J.H. 1991. Lithostratigraphic revision of the Upper Carboniferous to Lower Permian strata in the Cumberland Basin, Nova Scotia and the regional implications for the Maritimes Basin in Atlantic Canada. Bulletin of Canadian Petroleum Geology 39, 289-314.

Sanderson, D.J., Marchini, W.R.D. 1984. Transpression. Journal of Structural Geology 6, 449-458.

Schenk, P. E. 1967 The Macumber Formation of the Maritime Provinces, Canada; a Mississippian analogue to Recent strand-line carbonates of the Persian Gulf., Journal of Sedimentary Petrology 37, 365-376.1967.

---, 1969. Carbonate-sulphate-redbed facies and cyclic sedimentation of the Windsorian Stage (Middle Carboniferous), Maritime Provinces. Canadian Journal of Earth Sciences 6, 1037-1066.

---, 1970, Regional variation of the flysch-like Meguma Group (Lower Palaeozoic) of Nova Scotia, compared to recent sedimentation off the Scotian Shelf, in Lajoie, J., eds., Flysch Sedimentology in North America: Geological Association of Canada Special Paper 7, p. 127-153.

---, 1991, Events and sea-level changes on Gondwana's margin: The Meguma Zone (Cambrian to Devonian) of Nova Scotia, Canada: Geological Society of America Bulletin 103, 512-521.

---, von Bitter, P.H., Matsumoto, R., 1994, Deep-basin/deep-water carbonate-evaporite deposition of a saline giant; Loch Macumber (Visean), Atlantic Canada., Carbonates and Evaporites. 9, 187-210.

Stevenson, I.M. 1958. Truro Map Area, Colchester and Hants counties, Nova Scotia. Geological Survey of Canada, Memoir 297, 124.

St. Peter, C. 1993. Maritimes Basin evolution: Key geologic and seismic evidence from the Moncton subbasin of southern New Brunswick. Atlantic Geology 29, 233-270.

von Bitter, P.H., Scott, S.D., Schenk, P.E. 1992 Chemosynthesis; an alternate hypothesis for Carboniferous biotas in bryozoan/microbial mounds, Newfoundland, Canada. Palaios. 7, 466-484.

Waldron, J. W. F., 1992, The Goldenville-Halifax transition, Mahone Bay, Nova Scotia: relative sea-level change in the Meguma source terrane: Canadian Journal of Earth Sciences 29, 1091-1105.

---, 2004, Anatomy and evolution of a pull-apart basin, Stellarton, Nova Scotia: Geological Society of America Bulletin 116, 109-127.

---, and Jensen, L. R., 1985, Sedimentology of the Goldenville Formation, Eastern Shore, Nova Scotia: Geological Survey of Canada Paper, 85-15.

---, and Rygel, M. C., 2005, Role of evaporite withdrawal in the preservation of a unique coal-bearing succession: Pennsylvanian Joggins Formation, Nova Scotia: Geology 33, 337-340.

---, Piper, D. J. W., and Pe-Piper, G., 1988, Deformation of the Cape Chignecto Pluton, Cobequid Highlands, Nova Scotia: Thrusting at the Meguma-Avalon boundary: Atlantic Geology 25, 51-62.

---, Murphy, J.B., Melchin, M., and Davis, G., 1996, Silurian tectonics of western Avalonia: Strain-corrected subsidence history of the Arisaig Group, Nova Scotia: Journal of Geology 104, 677-694.

Webb, G. W., 1969, Paleozoic wrench faults in Canadian Appalachians, in Kay, M., eds., North Altantic Geology and Continental Drift: American Association of Petroleum Geologists, Memoir 12, 745-786.

White, S.H., Burrows. S.E., Carreras, J., Shaw, N.D., Humphreys, F.J. 1980. On mylonites in ductile shear zones. Journal of Structural Geology, 2, 175-187.

Williams, H., 1979, Appalachian Orogen in Canada: Canadian Journal of Earth Sciences, 16, 792-807.

---, 1995, Temporal and spatial divisions, in Williams, H., ed., Geology of the Appalachian-Caledonian Orogen in Canada and Greenland: Geological Survey of Canada, Geology of Canada No 6. Also Geological Society of America, Geology of North America, F-1, 23-44.

---, Colman-Sadd, S. P., and Swinden, H. S., 1988, Tectonic-stratigraphic subdivisions of central Newfoundland, Current Research Part 1B: Geological Survey of Canada Paper 88-1B, 91-98.

Withjack, M. O., Olsen, P. E., Schlische, R. W. 1995. Tectonic evolution of the Fundy rift basin, Canada: Evidence of extension and shortening during passive margin development. Tectonics 14, 290-405.

Yeo, G. M., Ruixiang, G. 1987. Stellarton graben: an upper carboniferous pull-apart basin in northern Nova Scotia. In Beaumont, C., Tankard, A.J. (Eds.), Sedimentary basins and basin-forming mechanisms. Canadian Society of Petroleum Geologists Memoir 12, 299-309.

Pre-conference Field Trips

A1 Contamination in the South Mountain Batholith and Port Mouton Pluton, southern Nova Scotia
D. Barrie Clarke and Saskia Erdmann

A2 Salt tectonics and sedimentation in western Cape Breton Island, Nova Scotia
Ian Davison and Chris Jauer

A3 Glaciation and landscapes of the Halifax region, Nova Scotia
Ralph Stea and John Gosse

A4 Structural geology and vein arrays of lode gold deposits, Meguma terrane, Nova Scotia
Rick Horne

A5 Facies heterogeneity in lacustrine basins: the transtensional Moncton Basin (Mississippian) and extensional Fundy Basin (Triassic-Jurassic), New Brunswick and Nova Scotia
David Keighley and David E. Brown

A6 Geological setting of intrusion-related gold mineralization in southwestern New Brunswick
Kathleen Thorne, Malcolm McLeod, Les Fyffe, and David Lentz

A7 The Triassic-Jurassic faunal and floral transition in the Fundy Basin, Nova Scotia
Paul Olsen, Jessica Whiteside, and Tim Fedak

Post-conference Field Trips

B1 Accretion of peri-Gondwanan terranes, northern mainland Nova Scotia and southern New Brunswick
Sandra Barr, Susan Johnson, Brendan Murphy, Georgia Pe-Piper, David Piper, and Chris White

B2 The Joggins Cliffs of Nova Scotia: Lyell & Co's "Coal Age Galapagos"
J.H. Calder, M.R. Gibling, and M.C. Rygel

B3 Geology and volcanology of the Jurassic North Mountain Basalt, southern Nova Scotia
Dan Kontak, Jarda Dostal, and John Greenough

B4 Stratigraphic setting of base-metal deposits in the Bathurst Mining Camp, New Brunswick
Steve McCutcheon, Jim Walker, Pierre Bernard, David Lentz, Warna Downey, and Sean McClenaghan

B5 Geology and environmental geochemistry of lode gold deposits in Nova Scotia
Paul Smith, Michael Parsons, and Terry Goodwin

B6 The macrotidal environment of the Minas Basin, Nova Scotia: sedimentology, morphology, and human impact
Ian Spooner, Andrew MacRae, and Danika van Proosdij

B7 Transpression and transtension along a continental transform fault: Minas Fault Zone, Nova Scotia
John W.F. Waldron, Joseph Clancy White, Elizabeth MacInnes, and Carlos G. Roselli

B8 New Brunswick Appalachian transect: bedrock and Quaternary geology of the Mount Carleton – Restigouche River area
Reginald A. Wilson, Michael A. Parkhill, and Jeffrey I. Carroll

B9 Gold metallogeny in the Newfoundland Appalachians
Andrew Kerr, Richard J. Wardle, Sean J. O'Brien, David W. Evans, and Gerald C. Squires