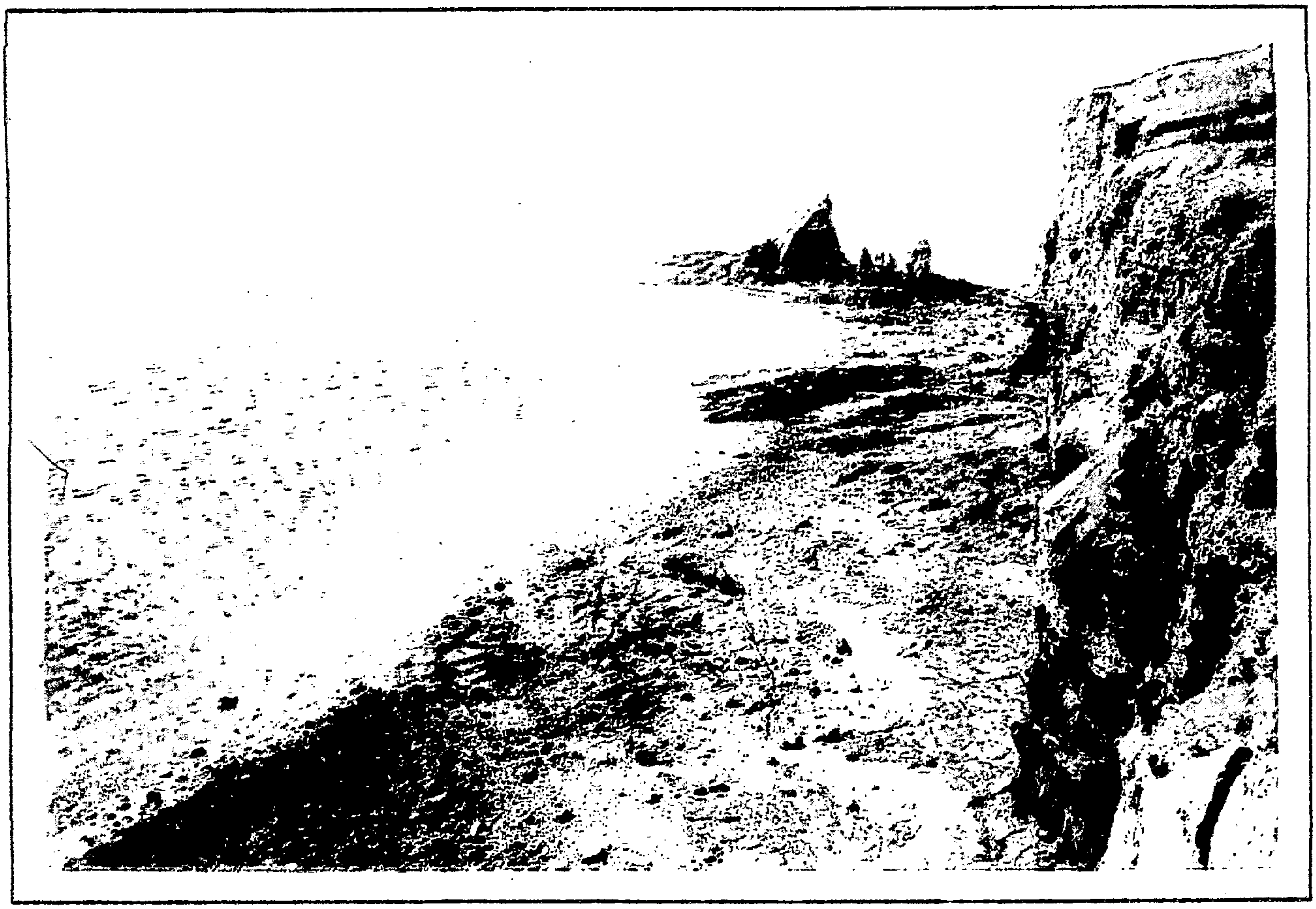
DISCOVERING ROCKS, MINERALS AND FOSSILS IN ATLANTIC CANADA

A GEOLOGY FIELD GUIDE TO SELECTED SITES IN NEWFOUNDLAND, NOVA SCOTIA, PRINCE EDWARD ISLAND, AND NEW BRUNSWICK



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Atlantic Geoscience Society
La Société Géoscientifique de L'Atlantique
AGS Special Publication 14

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I invite you to join the Atlantic Geoscience Society,
write c/o The Department of Earth Sciences, Dalhousie University (see above)

Cover Photo
Cape Split looking west into the Minas Channel, Nova Scotia. The split is caused by erosion along North–South faults cutting the Triassic-Jurassic-aged North Mountain Basalt and is the terminal point of a favoured hike of geologists and non-geologists alike. Photo courtesy of Rob Fensome, Biostratigrapher, Geological Survey of Canada (Atlantic), Bedford Institute of Oceanography.

CONTENTS

Introduction

This book offers exciting geological field trips to the whole of the Atlantic region. It is part of a growing sector in publishing that caters to people wishing to experience nature and natural phenomena. Local residents, as well as visitors from outside the region are taking up ecotourism; teachers are getting their students out of the classrooms, and professional guides are looking for alternative venues and material. Since the mid-1980s many guidebooks have found these audiences and this volume is one more addition to that growing library.

The idea for this book started with a suggestion by Graham Williams, after I had been asked for help on field trips by several undergraduate students and a couple of elementary school teachers. One of my jobs as Senior Instructor in the Department of Earth Sciences at Dalhousie University is to organise and coordinate most first-year trips, as well as to teach the Field Methods course. This gives me experience on places to visit within easy travelling distance of Halifax, as well as ideas on interesting places farther afield. I am also part of the organisation called Scientists in the Schools (SITS) and consequently I have frequent requests for help in the classroom and the field. In most situations, I refer to several series of geological conference guidebooks published by the Geological and Mineralogical Associations of Canada. However, these guides are for professionals and contain too much information, in the form of regional background material, as well as using complex language, and too many stops or outcrop descriptions for day or weekend trips. I usually end up copying the pertinent pages out of the guides and annotating them, drawing up detailed road logs on how to get to some of these places and much more. A single request sometimes takes hours. Writing a geological field guide for Atlantic Canada short circuits this process and makes an excellent 25th Anniversary Project of the Atlantic Geoscience Society (AGS).

The audience for this book has varied levels of expertise which makes for some complicated descriptions. There are teachers and students in the junior and senior public schools and professional geologists looking for a spot to do some weekend geology. As I am not an expert on the geology of Atlantic Canada, I solicited help from the Atlantic Geoscience Community to write some of the descriptions of their favourite areas. I wanted this book to be widely used, so field locations were sought around the provinces. I put a call out over the Internet and gave poster sessions at local geoscience society meetings, and I put the touch on friends and associates. The response was overwhelming, even with the limitations of the target audience. In the first few months I received more than 30 suggestions or written guides. In consultation with members of the AGS Education Committee, additional locations were chosen to fill in some gaps.

The first page of each location indicates its origin. If individuals wrote descriptions for this book I left them as authors but if the decriptions originated from a GAC/MAC fieldguide, I referenced the original. In this latter situation, I tried to use minimal interpretation and in many instances left wording similar to the original. Only rarely did I change ideas or data. Where no author is given, the description is my composition. As you can see from the list of authors, it is truly a collaborative effort.

Because of the multi-authored nature of this book the descriptions, style, maps, and format varies: I apologise to the reader at the outset for these inconsistencies. There is no way around this, although with editing and page design, some effort has been made to overcome this. Also, because of the varied levels of readership, some portions may seem unduly complex, while others seem over-simplified.

In the summer of 1996 I visited most of the sites listed here, to make sure the location maps and geology were accurate and the interpretations credible. In rare instances I changed descriptions and interpretations from the originals. While I can attest to the location and somewhat to the geological interpretation of most of the sites, it is left up to you, the reader and field person, to decide whether you agree with the interpretation. If you wish to enter into a discussion on the geology at any of these sites, you can either contact the original author or me, but I warn you I do not have the definitive answer (nor, I suspect, does the original author).

To the many individuals who helped in this book I extend a warm gesture of gratitude. These are all the authors listed but also individuals such as Alan Ruffman, Graham Williams, John Waldron, Steve McCutcheon, Jim Walker, Charlie Walls, Tom Duffett, Jeremy Wallace, and Carolyn Green get a special mention. Without all your help this would never have gotten off the ground.

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The following individuals deserve special mention as contributors and writers of some of the field desciptions:

Claudia Riveros, William R. Jamieson, and **Peter A. Cawood** — Brigus: Precambrian deformation and the Precambrian-Cambrian Unconformity Boundary

Rebecca Jamieson — St. Anthony: Metamorphism Beneath and Ophiolite

John Waldron — Black Cove and Black Point: Traversing from the Shelf Succession to the Humber Arm Allochtohon
—Rainy Cove: The Carboniferous-Triassic Unconformity
— Big Tancook Island: The Rocks of the Goldenville-Halifax Transition
— Peggy's Cove: A piece of Nova Scotia

Tammy Allen — Port Hood: The Upper Carboniferous Colindale Member Type Section

Dan Bourque — Georgeville: Early Cambrian Intrusive Rocks
— Arisaig Pier: Rocks of the Ordovician Aged Bear Brook Volcanic Group

H. V. (Don) Donohoe — Victoria Park: Understanding the Importance of Rivers

David Piper and **Georgia Pe-Piper** — Spicer Cove and Squally Point: Volcanic Rocks, Dikes, and Thrust Faults

John Calder — Joggins: The Fossil Cliffs of Joggins: One of the World's Classic Geological Sites

Sandra Barr — Ross Creek Brook: A Section Through a 200 Million Year Old Baslatic Lava Flow

Sandra Barr and **Chris White** — Taylor's Head: Geology of the Irving Nature Park

Rick Horne — The Ovens: Gold Bearing Veins and Their Relationship to the Structural Geology of the Meguma Group

G.C. Milligan — Cedar Dunes Provincial Park: Beaches

Nina Chrzanowski and **John Spray** — Fredericton: Bedrock geology of the City of Fredericton

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W.D. Boyce	P.M. Myrow
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Safety and Ethics on Field Trips

Introduction

Geology as part of the Mining Profession is an inherently dangerous occupation. Consequently, going on geology field trips is dangerous and should not be approached with a cavalier attitude. Safety should be paramount in all situations and is a head space that must exist in our minds before true safety can be obtained. Safety implies avoidance, and accident prevention implies anticipation. It is through the process of awareness that we can achieve, avoid and anticipate. With most occupations the workplace is custom designed and built with safety hazards engineered out *(hopefully)* but with outcrops and geography, conditions are not engineered, they are constantly changing and are inherently dangerous.

Safety awareness can be raised by doing a few simple things. First, before going on any trip carry out an operations safety-risk-hazard assessment. Think about what is known about possible risks and causes through reviews of past trips and their problems. Read what is written about safety at the beginning of each description in this book before going on the trip. Second, assess the equipment that you are using including your vehicles as well as the geological equipment such as compasses and hammers. Third, assess the regulations, guidelines, and procedures that go with this equipment. And fourth, evaluate your knowledge and that of the others accompanying you on the above and take corrective measures before you start. Basically most accidents result from performing an unsafe act, tolerating an unsafe condition or hazard, or failing to follow proper procedures.

General responsibilities

The general responsibilities for everyone on any field trip are few but must be stated. First all machinery (cars and trucks getting you to the outcrops) should be in good working order and the operation of the same must be done in a safe manner. Second, everyone on a geology field trip should be able to read a map and do basic orienteering. This should hold for anyone going on a hike into the woods or along the coast. Third, the leader of the trip (and there should be one) should impart to everyone not to take risks and the rest of the group should import to the leader that they will not take risks. Fourth, everyone should know the chain of responsibility and their individual responsibilities in case someone does get into trouble. Most people go into a state of emotional shock when faced with an emergency but this can be overcome by prior training and talking over the safety issues.

No matter what type of field trip you are on, whether a class trip or a group of individuals getting together for the weekend, everyone has a legal responsibility based on law. It is practically impossible to accurately predict any legal consequence but there are a few broad legal contexts that must be born in mind. In Contract Law the organiser of a field trip has a contract with the participants regardless of money transfer or written documents and in Tort Law the organisers of a trip (or anyone in a position of authority or expertise) may owe a greater duty to those who rely on them than ordinary individuals do. Torts also imply that an organiser of a field trip or field crew has more responsibility toward a novice field person than an experienced one so the onus is not only on the leader but also partly on the participants. Basically make no representations about services that are not directly under your control, i.e., the motel or campsites, quality of the outcrops, etc. Organisers of some trips may attempt to circumvent this by issuing waivers but all waivers can be gotten around by smart lawyers so you can sign them or not, it is up to you. Just make sure you know what you are signing. Remember, courts always have the last word and they also have the advantage of hindsight so be responsible.

Precautionary Behaviour

The following list is by no means exhaustive. It is what I generally tell my students prior to trips but I also recognise that some sections or trips need special consideration. Think about where you are going before the trip and plan beforehand.

Roadcuts

Avoid crossing highways to examine roadcuts; examine the rocks only on the side on which you are parked. On paved highways pull right off the road, do not park on the hard shoulders and do not allow participants to walk on the pavement. At all roadcuts walk, stand, take pictures, etc. off the road bed even if it means standing near a wet ditch, do not stray onto the road. Most rocks in a roadcut are highly fractured, loose, and unsteady. Do not climb the cuts, stay back from the rock face, do not hammer above your waist, and do not go under overhangs.

Cliff sections

Falling rocks are major hazards at cliff sections whether along roadcuts or coastal exposures. Avoid obviously unstable cliffs as well as overhangs, do not hammer above your waist, wear hard hats at all times and do not climb the cliff face. Plan slope-bottom activities away from the base of the cliff. A good rule of thumb is to stay away a distance equal to the height of the cliff. If you must climb a steep section do not allow participants to climb while others are below and do not allow participants to go near to the top of the slope, stay well back. Do not plan slope-bottom and slope-top activities at the same time. Do not throw rocks down slopes, or throw them off the tops of cliffs. Do not run down hills or slopes.

Tides

Take great care on a section that is cut off at a high tide. Do not work on a rising tide. Tidal sections are usually wet and muddy with lots of flora and fauna so as a consequence, slippery and dangerous. Do not jump around from pinnacle to pinnacle and wear appropriate footwear. If working with a group make sure someone is aware of the state of the tide at all times and brings up the rear of the party to watch for stranglers.

Old mines

Do not enter old mine workings.

Working mines

Do not enter working mines except by arrangement, and always in company. Use appropriate clothing and make sure someone knows where you are, when you went underground, and when you are expected back. Make sure there is adequate ventilation. Do not pick up explosives or any equipment laying about. Do not pull at wires. Meet the requirements with regards to parking, trespassing etc.

Quarries and Pits

The same can be said for these as for mines, do not enter without permission and use proper care.

Clothing

At all times have on adequate footwear. Wear layers of clothing that will cut the wind and be gradually shed as the day warms. Hats are important for sun protection as are gloves to protect your hands. Good clothing is important if there should be an accident.

First Aid and safety

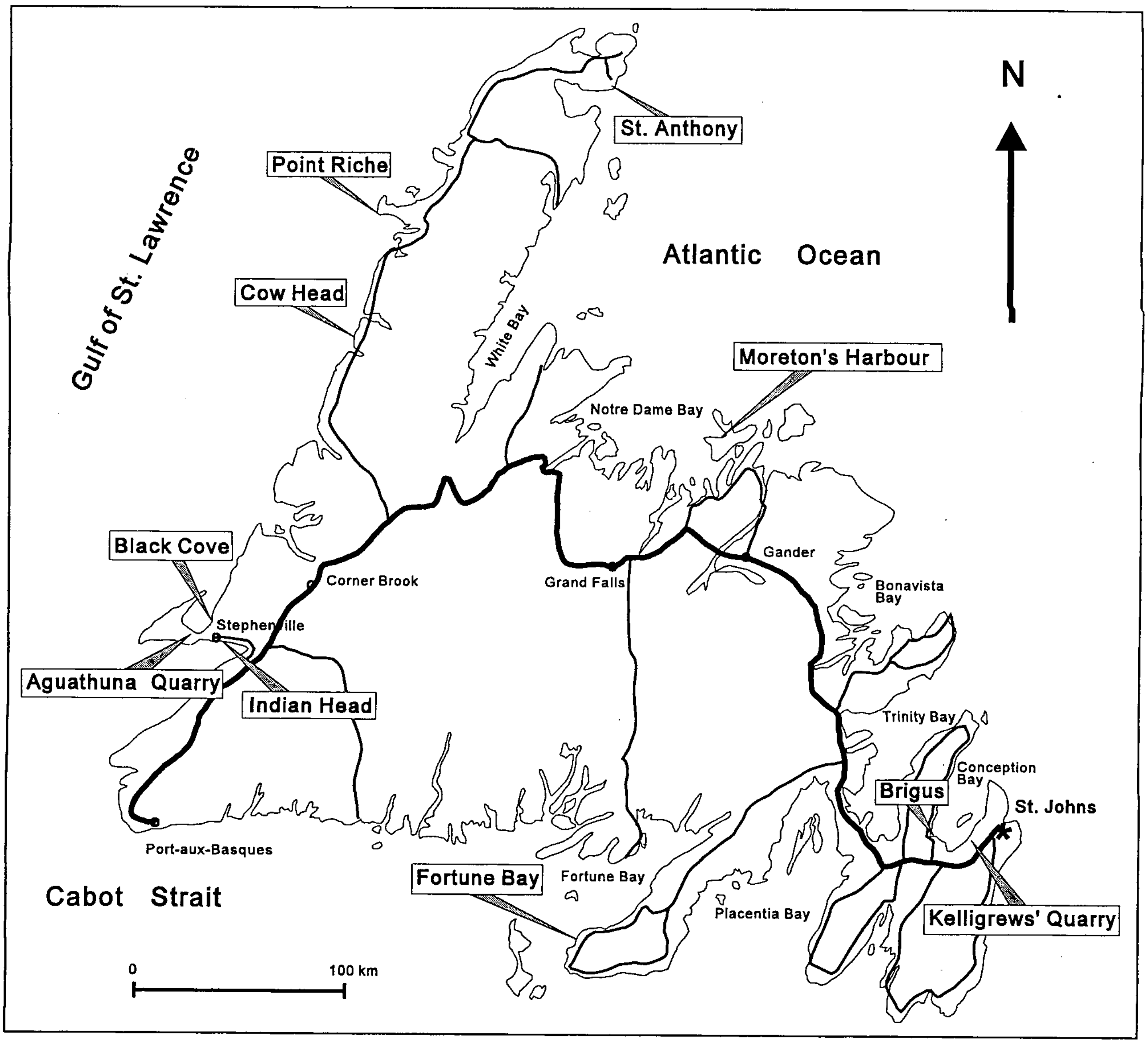
Everyone on the party should have some type of first aid kit, some high-energy food, and a whistle. Carry a compass, matches in a waterproof container and do field trips in groups. Let someone know where you have gone, the route you plan to follow, and when you expect to be back. Take no unnecessary risks.

Ethics

The literature abounds with examples of key outcrops and specimens being destroyed by weekend amateurs as well as professional geologists, quarries and roadcuts declared off limits because of liability, and private property closed to visitors because of vandalism. I also believe that if we did not have collectors many of our fine mineral specimens, rock samples, and fossils would not be star exhibits in our museums, private collections, schools, or universities; they would have been destroyed by the elements. The problem is to balance between these two extremes.

What can be done to minimise our effect on the environment? What are your responsibilities? There is no one answer to these problems but a discussion on them will go a long way. First is the basic ethics of going along the coast or road on a hike. Think about what is appropriate behaviour with regards to activites, garbage, and private property. Be discrete with regards to your toilet activities. Pack out what you bring in and leave the area neat and tidy. I had a very good friend who always took along an extra garbage bag to pack out previous user's garbage as well as his own. We could all try to imitate him. Second is the ethics at the outcrop. The coast is a very actively eroding area as are most roadcuts but it is limited and damage done to an outcrop takes time to heal to look "natural" again. Do not sample mindlessly, hammer with discretion, and leave hard to get at samples for another day. Do not deface, sample or otherwise destroy key geological phenomena. When possible collect only the loose debris found on the beach. Remember you are here only a lifetime but rocks are here forever.

NEWFOUNDLAND



Outline map of Newfoundland showing locations of field trip descriptions

Kelligrews Quarry

Trilobites of the Manuels River Formation

Purpose

To collect and identify trilobites.

Location

Take Highway 2 to Manuels and then south on Highway 60 (Fig. 1). Turn left onto the Red Bridge Road about 7 km from the intersection of Highways 2 and 60. Walsh's Road intersects the Red Bridge Road about 1.3 km from the highway and is the location of the first small disused quarry. The other quarries stretch along behind the trees and boulders on the left for about 0.6 km.

Introduction

The Cambrian trilobite faunas of eastern Newfoundland differ from those of western Newfoundland and the rest of North America because the two areas occupied different climatic and geographic zones on the opposite sides of the Iapetus Ocean. During the Cambrian, western Newfoundland (and most of North America) was situated near the equator in relatively warm, shallow water, whereas eastern Newfoundland was located in higher latitudes in a temperate to polar climate and was covered by relatively cool, deep water. The fossil Cambrian faunas of eastern Newfoundland are collectively part of the Acado-Baltic faunal province (named after the Acadian provinces of New Brunswick, Nova Scotia and Massachusetts and the Baltic countries of Norway, Sweden, England, Wales, France, Spain, Germany and Bohemia) and are some of the most complete successions in the world.

When looking for trilobites, it is important to remember that complete specimens are rarely found. This is because trilobites grew by periodically shedding their exoskeletons, as crabs and lobsters do today. Most trilobite remains are therefore discarded bits of shell rather than the complete animal. While complete specimens are better for scientific description in paleontological studies, even a poorly preserved head or tail is often enough for field identification.

Trilobites can be found by picking through weathered rubble whereas finding them in place requires a careful layer-by-layer examination of the enclosing sedimentary rocks with a hammer and chisel. Many specimens are quite small and a magnifying glass or hand lens is quite handy for checking specific features or the rock in general. Whenever using a hammer and chisel always use safety glasses.

Broken fossil specimens can be field repaired using nontoxic white glues while nonretrievable specimens can be reproduced using plaster, modelling clay, or a liquid latex. Fossils which cannot be collected may also be photographed or sketched. After collection, all specimens ought to be securely wrapped in tissue or newspaper and then placed in a labelled bag to prevent damage during transportation. It is also a good idea to note the location of the fossil collection on a map or in a field book in order to make it easier to find again if the need arises.

Fossil collecting is against the law in National and Provincial Parks and Ecological Reserves, except with a special permit.

Trilobite Morphology

Trilobites are an extinct class of marine arthropods that inhabited the earth's oceans from the Early Cambrian Period to the Late Permian Period, a time span of approximately 300 million years. They are so named because they are divisible along their lengths into three lobes - a central axial lobe and two pleural lobes. Trilobites are also divisible into three parts front to back - the head (cephalon), the body (thorax) and the tail (pygidium).

Like modern crustaceans (crabs, lobsters, etc.) arachnids (spiders, scorpions, etc.) and insects (flies, mosquitoes, etc.) trilobites had compound eyes made up of very closely packed individual lenses, which gave them a wide field of view. Some trilobites were secondarily blind, however, having no eyes developed. Trilobites also had sensory antennae.

The trilobite shell or exoskeleton was composed of calcium carbonate and chitin; the later comprises hard, shiny, flexible layers of fibrous, nitrogen-rich polysaccharide — a hydrocarbon related to cellulose. Like modern crabs and lobsters, trilobites had to periodically shed their exoskeletons, or molt, in order to grow. Figure 2 labels the most important parts of the trilobite exoskeleton and can be compared with drawings of actual speciments, as in Figure 3.

Trilobites lived in a wide variety of marine environments, such as i) very shallow water and nearshore tidal flats, ii) offshore sand bars, iii) lagoons and reefs in inland seas, at edges of continental shelves, and around volcanic islands, and iv) deep water off the edges of continental shelves. Trilobites crawled or swam along the sea floor or burrowed into it. This activity produced fossil tracks, trails, and burrows. Some trilobites actively swam or passively floated higher up in the water column. When attacked, some trilobites rolled up into protective balls. Others had extremely thick, strongly curved shells for protection against predators or to withstand buffeting against rocks by strong currents. Some had numerous sharp spines or wide fringes both for protection and for support on muddy sea bottoms. No one knows how or what trilobites ate or how they mated, although there is no lack of educated guessing.

Trilobites are particularly useful for dating rocks of the Cambrian and Ordovician systems. Trilobites had a high rate of evolution during the Cambrian and Ordovician periods. The composition of the faunas changed many times so that many trilobite zones are recognized. At the same time, trilobites were abundant and widespread. This has made it possible to correlate trilobite faunas of the same age over great distances.

Kelligrews Quarries

The Middle Cambrian Chamberlain's Brook Formation (Adeyton Group) and Manuels River Formation (Harcourt Group) are well exposed here. The Chamberlain's Brook Formation is exposed in the large actively used quarry. A 15–40 cm thick distinctive orange-brown weathering, nodular "rottenstone" trilobite-coquina limestone forms a useful marker within the formation. The bed is known as a coquina because it is packed with abundant shelly remains, in this case trilobites, which were presumably concentrated by current action along the shoreline of the shallow Cambrian sea. Generally, the beds below the "rottenstone" marker are sparsely fossiliferous, whilst those above are abundantly fossiliferous. The "rottenstone" marker horizon probably corresponds to beds 21–25 of Howell's (1925) Manuels River section, which occur 71.41–72.06 m above the base of the Chamberlain's Brook Formation.

A transitional zone of dark blue-grey mudstone and shale separates the Chamberlain's Brook Formation from the conformably overlying Manuels River Formation. At Manuels River, the boundary is defined by a thin (3 cm) layer of recessive weathering volcanic ash, informally known as the Manuels metabentonite. An apparently analogous horizon also occurs in the upper part of the large quarry.

The Manuels River Formation is exposed in the uppermost part of the large quarry and in the small abandoned quarry at the intersection of the Red Bridge Road and Walsh's Road. Fossils are locally abundant.

In the large quarry, above the "rottenstone" marker and below the Manuels River Formation are recorded 14 different species of trilobites in the single trilobite zone present. The Manuels River Formation in the same quarry as well as the small quarry has yielded 19 different species in 2 trilobite zones.

Source

Boyce, W.D., 1988. Cambrian Trilobite Faunas on the Avalon Peninsula, Newfoundland. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, St. Johns '88; Field Trip A8, 52 p.

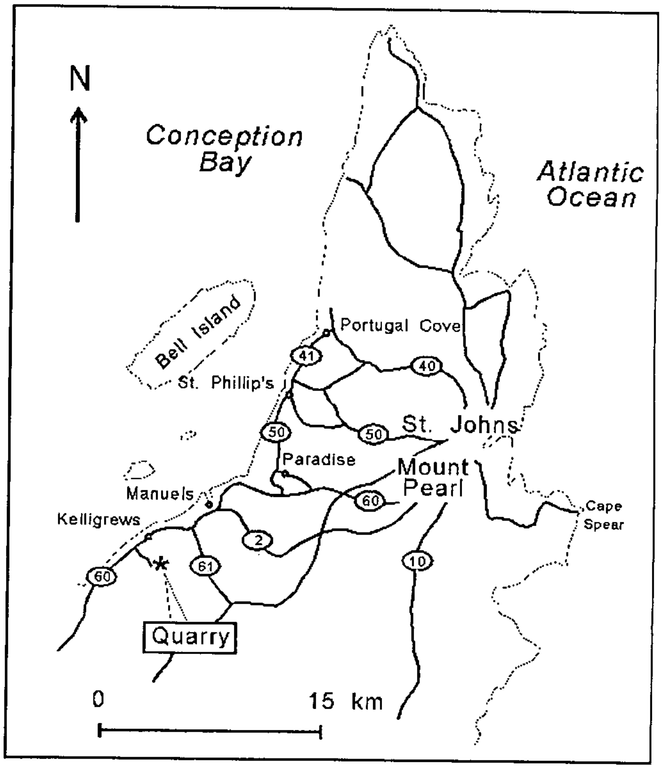


Figure 1: Location map of Kelligrew's Quarry

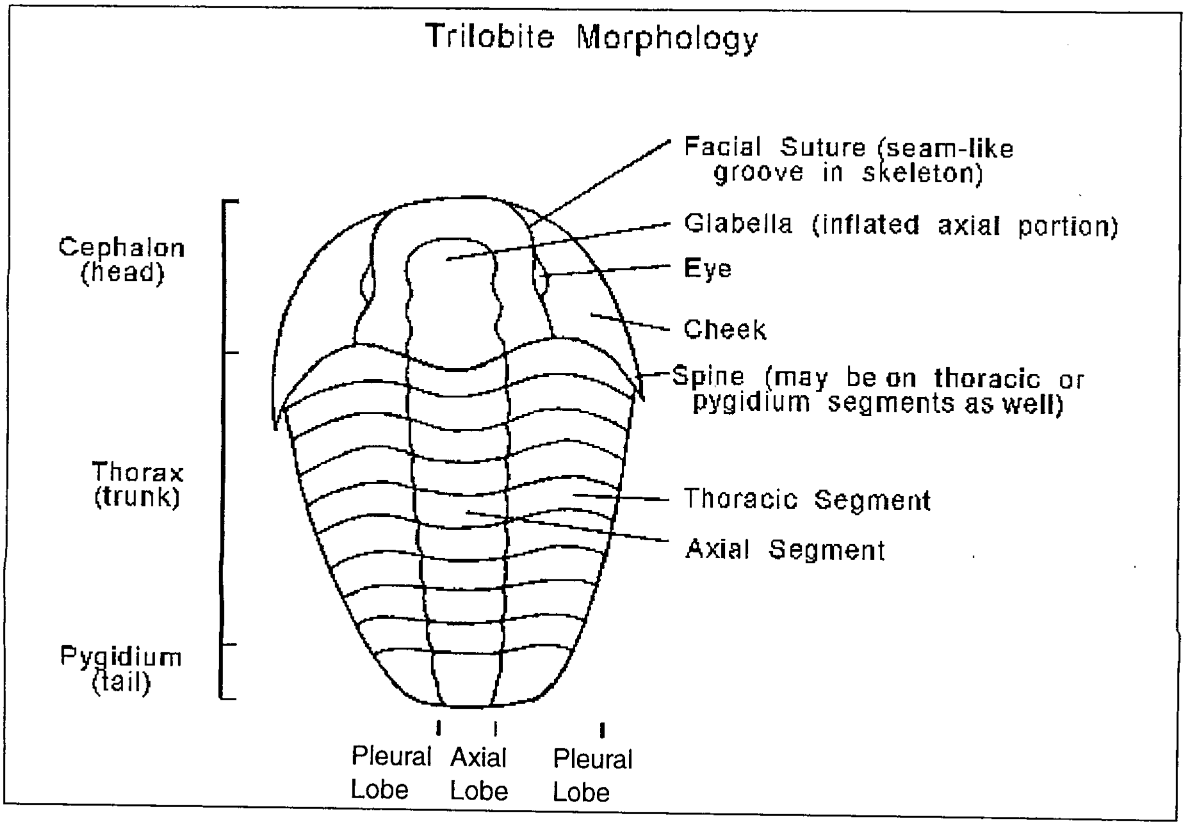


Figure 2: Sketch showing the major parts of a sample Trilobite

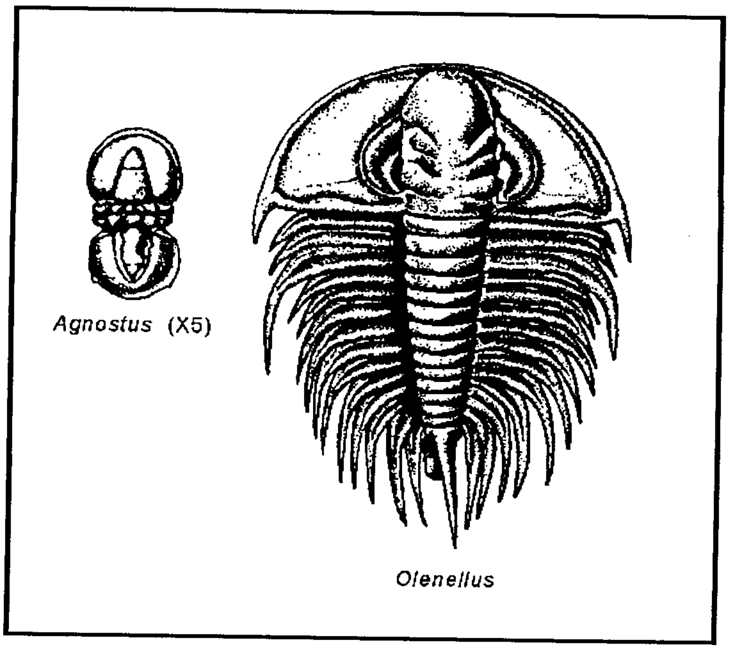


Figure 3: Two examples of trilobites from the Cambrian Period. Compare with Figure 2. Agnostus is found in the Manuels River Formation whereas Olenellus is not: it is shown only to demonstrate a species with spines plus it is the most common example of a trilobite illustrated in textbooks

Brigus

Precambrian Deformation and the Precambrian-Cambrian Unconformity Boundary

Location

The Brigus area is located along the southwestern shoreline of Conception Bay, about 2 km southeast of the very picturesque community of Brigus, a 1.5 hour drive from St. John's (Fig. 1). From St. John's along the Trans Canada Highway west, exit north on the Avondale access road. At the Avondale junction take Highway 60 and continue west to Brigus.

Introduction

The Brigus area is of geological significance because it provides unequivocal evidence for Precambrian deformation. Elsewhere in the Brigus area, basal Cambrian units overlie massive Harbour Main Group volcanic rocks, which do not display structures that so clearly establish a temporal control on Precambrian deformation. The field trip entails four stops. Exposed at the fourth stop, Seal Head, are Precambrian rocks of the Harbour Main Group that are overlain by Adeyton Group Cambrian rocks with an angular unconformity. This surface indicates that Precambrian rocks were significantly folded and faulted during Precambrian deformation, prior to the overlap of basal Cambrian rocks. Thus, the angular unconformity surface provides a temporal control on the structures in the underlying Precambrian rocks.

Regional Geology of Newfoundland

The Appalachian Orogen in Newfoundland is part of the Paleozoic Appalachian-Caledonian mountain chain. Before the Mesozoic opening of the Atlantic Ocean, this chain extended continuously about 7 500 km from the Caledonian Orogen in east Greenland-Scandinavia in the north, south to the Appalachian Orogen in the eastern United States and the Mauritanian Orogen in west Africa (Williams 1976). The Appalachian Orogen in Newfoundland is divided into four major tectonic-stratigraphic zones which include the Humber, Dunnage, Gander, and Avalon zone (Fig. 2).

The Avalon zone collided with the Gander zone during the middle Paleozoic (Devonian) Acadian Orogeny (Dallmeyer et al. 1983), consequently deforming rocks in the Avalon zone. Prior to the Acadian Orogeny, the Avalon zone is thought to have been deformed by an older orogenic event during the late Precambrian known as the Avalonian Orogeny (Lilly 1966). The focus of this field trip is to present controls on the style and relative age of the deformation associated with the two orogenic events in the Avalon zone. The style is expressed in terms of structural elements (e.g., folds, cleavage, faults), while the relative age of deformation is controlled by cross-cutting relationships of structures (e.g., by the angular unconformity surface).

Geology and Structure of the Brigus Area

Rocks in the Brigus area (Fig. 3) are part of the Avalon zone and are relatively unmetamorphosed. Late Precambrian rocks include the Harbour Main Group, comprising dominantly volcanic sequences with minor sedimentary units, and the younger Conception Group comprising a deep marine turbidite sedimentary sequence. In this area, the contact between Harbour Main and Conception Group rocks is the steeply west dipping Brigus fault zone, displaying a braided morphology. Lower Paleozoic marine shelf sedimentary rocks include the lower and middle Cambrian Adeyton and Harcourt Groups, respectively. The base of the Harcourt Group is defined by the occurrence of manganiferous beds, and is not further discussed in the field guide. The contact between Lower Paleozoic and late Precambrian rocks is unconformable.

Rocks of the late Precambrian Harbour Main and Conception Groups record structures formed during both the late Precambrian (Avalonian) and middle Paleozoic (Acadian) orogenic events, while lower Paleozoic rocks record structures formed during the middle Paleozoic (Acadian) orogenic event. Crosscutting relationships with the regional (S2) cleavage that form in Cambrian rocks are an important tool in determining the relative age of structures that formed in both Precambrian and Cambrian rocks. The pervasive cleavage formed in all lithologies across the Brigus area is related to the second period of deformation (Acadian) and will therefore be referred to as S2 cleavage.

Excursion Stops

The excursion entails an approximately 2 km long, interesting, and scenic hike to Seal Head (Stop 4) (Fig. 3). Parking is available at the beginning of the trail, just west of Frog marsh. Each stop is described in terms of location and a brief description of field observations and relationships.

Equipment needed: a comfortable pair of hiking or rubber boots for the relatively long hike, possibly muddy areas along the trail and sharp and sometimes slippery rocks; a warm jacket for commonly strong winds along the shoreline and binoculars if possible.

Caution: the beginning of the trail passes over private property which should be respected. Avoid walking on talus slopes which locally form along cliffs that reach up to 60 meters in height in the Seal Head area. DO NOT attempt to climb down to the beach.

Stop 1 Conception Group rocks

Location: By the cobble beach along the eastern shore of Lobster cove.

Description

Here, moderately northwest-dipping pale greenish-grey siltstone and pelite interlayers of the Conception Group show strong bedding, however the regional S2 cleavage is not well developed. Notice the large amounts of quartz veining in the rocks, probably representing a narrow deformation zone along the Lobster fault. Veins are brittle structures which form in dilational and/or shear fractures in rocks. Fluids (in this case rich in silica) are flushed through the rocks and fractures subsequently precipitating silica (quartz) in the fractures. Looking across Frog marsh, the Brigus fault can be seen along the gully that separates the marsh (underlain by Conception Group rocks) and the ridge comprising Harbour Main Group red tuffs.

The first 200 m of the excursion trail follows the trend of the Lobster fault (Fig. 3). As you head towards Stop 2, note the small log bridge and passing stream. The stream occurs in Conception Group rocks and delineates the Brigus fault (Fig. 3). As you make your way eastward along the trail from the bridge to the top of Red Rock Pond, you will encounter outcrops of Harbour Main Group rocks, mainly of red and grey massive volcanic tuffs and pyroclastic breccias (which do not show any obvious layering). S2 cleavage is commonly weak or absent in these rocks.

Stop 2 Lower Cambrian rocks

Location: Stop 2 occurs adjacent the eastern shoreline of Red Rock pond and is approximately 1.1 km from Stop 1. Along the way, examine the blocks of Cambrian red pelites and pink limestone and nodular limestone beds that have fallen from the ridge to the east.

Description

S2 cleavage is commonly well developed in red Cambrian pelites and less well developed in pink Cambrian limestone and nodular limestone beds, with significant S2 refraction (change in orientation—Fig. 4 dii) across limestone-pelite boundaries. In limestone, beds buckle and form cm-scale open folds (F2) and associated axial planar S2 cleavage (fig. D di). F2 cm-scale folds do not develop in the pelites.

The intensity of S2 cleavage is a function of mechanical rock properties, which in turn is a function of rock strength. Stronger limestones are not as easily deformed as are the weaker pelites. Cleavage refraction can also be accounted for by a difference in competency between the different lithologies (Fig. 4 dii). Amid the talus, one can find hackle-mark structures on fracture surfaces (Fig. 4 ei) as well as pencil cleavaged fragments (Fig. 4 eii). These fragments, as the name suggests, are pencil shaped and vary in size.

Stop 3 Harbour Main Group conglomerates

Location: Stop 3 occurs about 160 m south of Stop 2. Conglomerates at Stop 3 outcrop on a small land mass projecting into the pond, just west of the trail.

Description

Greenish-grey silicified conglomerates of the Harbour Main Group contain up to cobble size clasts with compositions ranging from rhyolitic to basaltic. Conglomerates are part of the sedimentary belt of the Harbour Main Group in the area.

Stop 4 Sub-Cambrian angular unconformity

Location: Stop 4 is approximately 530 m from Stop 3. Once you have reached the stream that pours into Small cove adjacent the pond, carefully make your way down to Seal Head (Stop 4) to get a good sectional view of the rocks on the west and west-northwest side of Small cove. Unfortunately, there is no obvious trail to follow down to Seal Head from the stream, however, the intervening tree stand is relatively sparse and manageable to walk through. Be sure to scan the water for minkies and humpbacks!

Description

This stop provides a spectacular sectional and map view of the angular unconformity surface (Fig. 4 b) between gently east-dipping lower Cambrian beds and steeply southwest-dipping Harbour Main Group sedimentary layers (Fig. 5).

Harbour Main Group sedimentary units comprise mainly sandstones with red tuff interlayers. Harbour Main Group tuffs are also in fault contact with the Harbour Main Group sedimentary sequence along the east-directed Flounder reverse fault (Figs. 4 fii; 5). In the footwall of the Flounder fault are two east-vergent slightly overturned anticlines (Fig. 4 c) in layered Harbour Main Group sequence. The folds plunge moderately to the south-southwest and diminish in amplitude, becoming more open in style (Fig. 4 a) upsection (from right to left). The folds eventually die out completely at the contact with the conformable 5 m thick red tuff layer. These folds are probably associated with reverse displacement on the Flounder fault. The Flounder fault and folded Harbour Main Group beds are cross-cut by and terminate against the sub-Cambrian unconformity surface, which clearly establishes these folds (F1) and fault to be Precambrian structures. On the way to Stop 4, you crossed over the Boot fault (Fig. 3). *The Boot fault* steeply dips to the east-northeast and is a brittle zone up to 3 m wide of quartz and tectonic breccia. Slickensides on fault surfaces display dip slip striae (Fig. 4 fi). Based on the fault striae orientations and the relatively wide brittle deformational nature of the zone, the fault probably underwent normal displacement in which the footwall (and Stop 4) was displaced downwards relative to the hangingwall (Fig. 4 fi).

The tight F1 folds beneath the unconformity surface locally develop axial planar cleavage in the fold hinge zone (which you probably will be unable to see from stop 4). This cleavage is termed S1. The more pervasively developed S2 cleavage (clearly formed in Cambrian beds) cross-cuts S1 and their fold axial surfaces by up to 20° in map view. In profile view, S2 is sub-parallel to S1.

A brief summary of what you have seen in stop 4 includes Precambrian structures in the sedimentary sequence of the Harbour Main Group comprising overturned, tight to open folds (F1) and developed axial planar S1 cleavage, associated with east-directed reverse displacement on the Flounder fault. These structures imply that the first deformational event (Avalonian) included at least a folding phase in the area. A temporal constraint on these structures is provided by the unconformable overlap of basal Cambrian units on folded Harbour Main Group beds, as well as their superimposition by S2.

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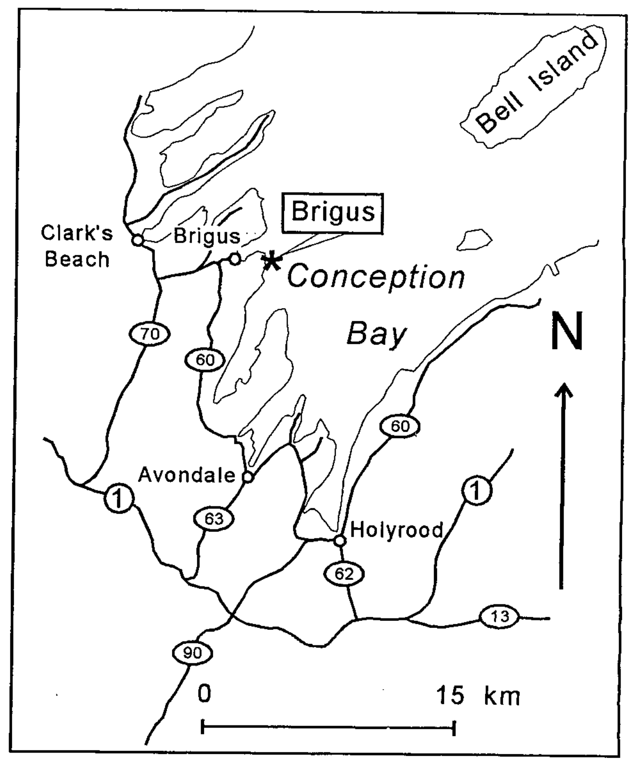

Figure 1: Location map of the Brigus Area, Conception Bay.

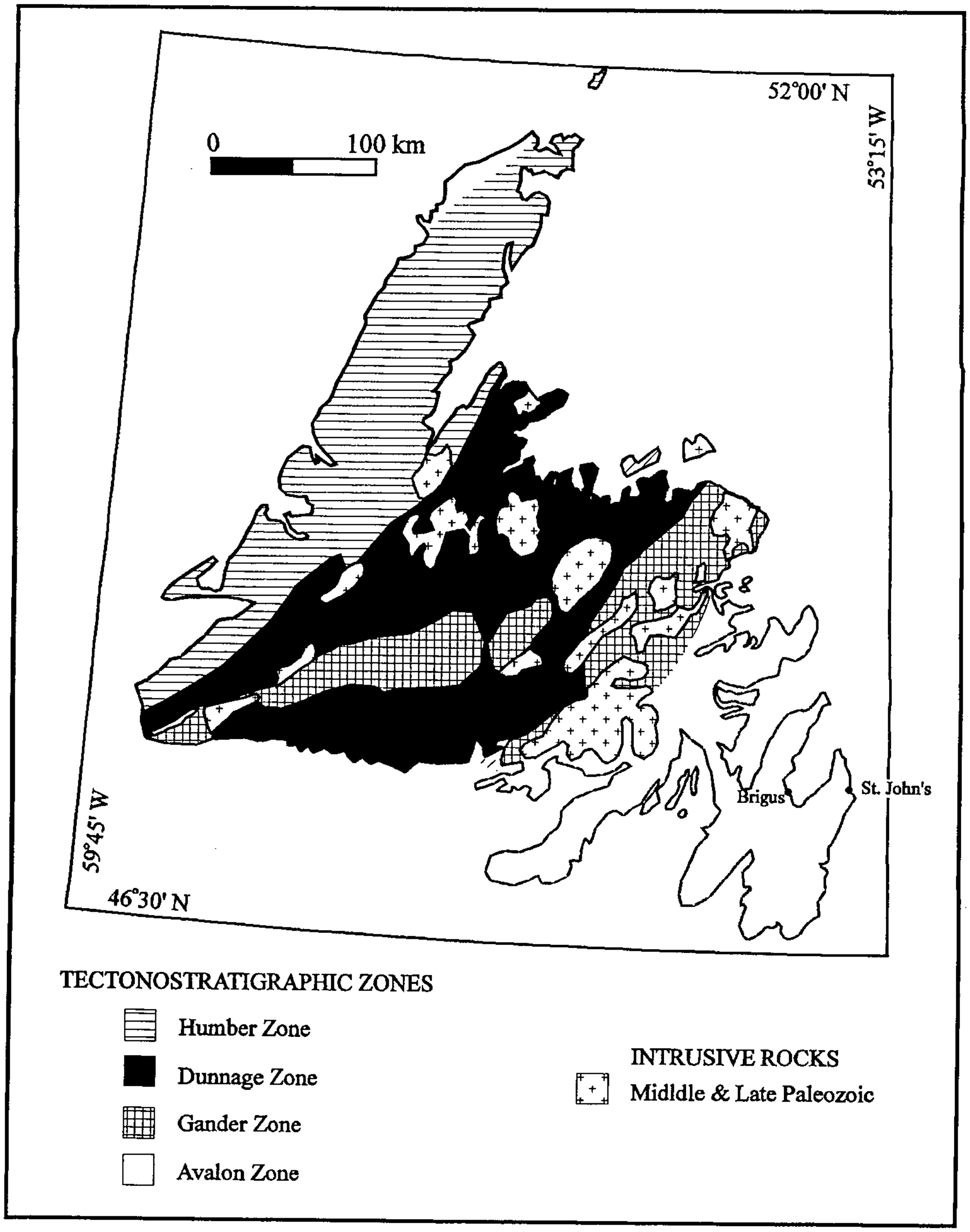

Figure 2: Tectonostratigraphic zones and instrusive rocks in the Newfoundland Appalachian Orogen (Williams, 1976)

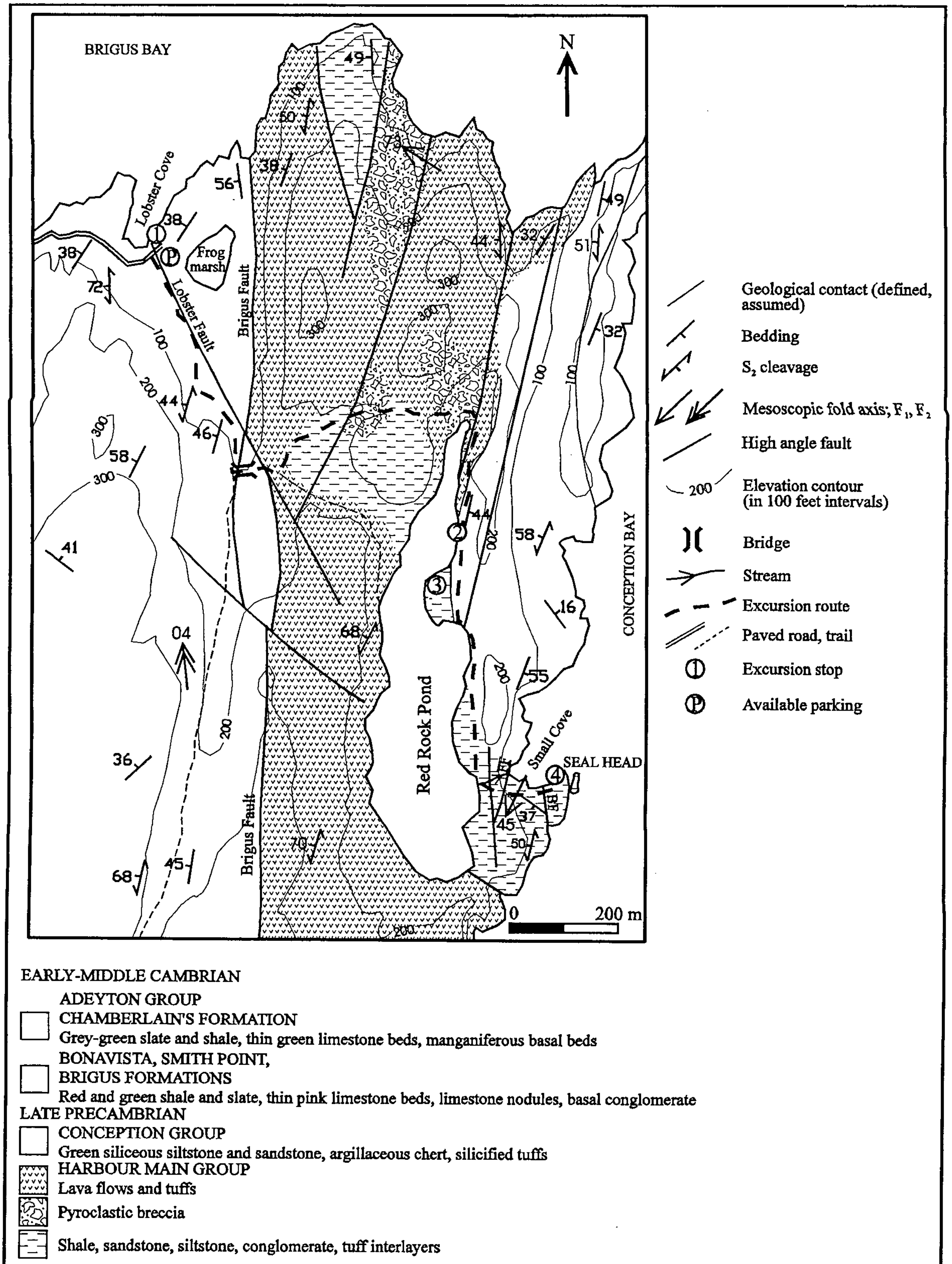

Figure 3: Detailed structural map of the Brigus area showing the excursion route and stop locations. BF- Boot Fault, FF- Flounder Fault.

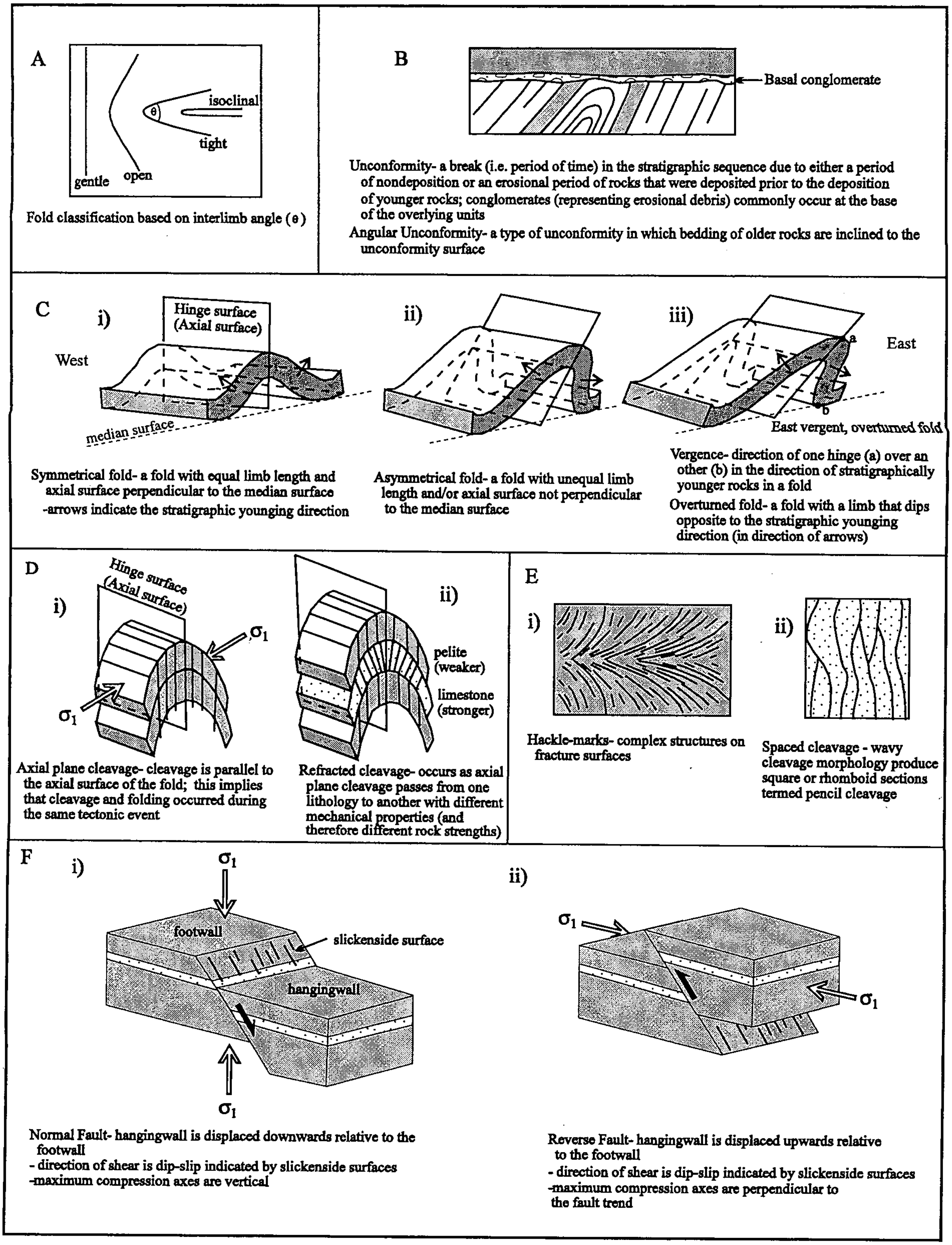

Figure 4: Summary sketches of some primary and secondary structures (modified from Price & Cosgrove 1990 and Hobbs *et al.* 1976)

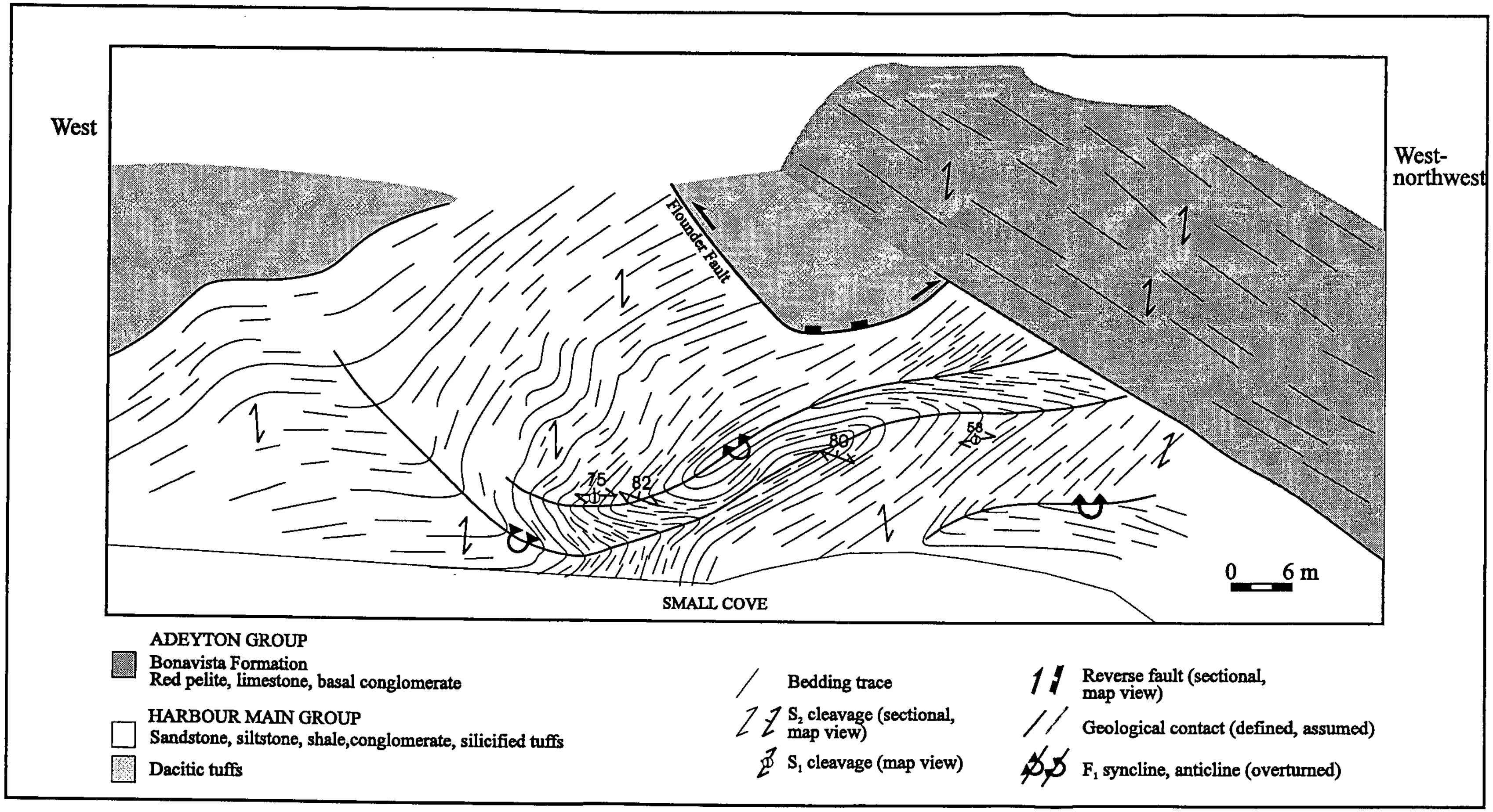

Figure 5 Stop 4. Sketch of the sub-Cambrian angular unconformity where Cambrian rocks overlie Harbour Main Group sedimentary and tuff interlayers. Directly underlying the unconformable surface, Harbour Main Group sedimentary rocks are folded into tight, overturned, east-vergent F1 folds, locally showing axial planar S1 cleavage. Up-section (to the left), the folds become more open, diminishing in amplitude until they eventually die out at the contact with the red tuff interlayer. The east-directed Flounder fault accommodates the folding, displacing red tuffs over the folded sedimentary units. S2 cleavate is weak in Harbour Main Group rocks and cross-cuts F1, fold axial surfaces and S1 cleavage in map view by up to 20°. In sectional view, S2 is sub-parallel to S1. Note that the sketch provides both map and sectional views.

Fortune Bay

The Precambrian-Cambrian Boundary Type Section

Purpose

The Chapel Island Formation at Fortune Bay exposes a superb section of the PreCambrian-Cambrian boundary that contains world class trace fossils.

Location

Fortune Head lighthouse is reached by a gravel road 2.7 km south of the Fortune town centre on Route 220 (Fig. 1). Park at the fork in the road near the lighthouse. The base of the Fortune Dump section lies several hundred meters east of the lighthouse and the section extends west along the coast for approximately 1 km. Descend a small gully on the east side of a promontory located just east of a small cove that acts as the town dump. Shallowly-dipping redbeds of member 1 are cut by a minor fault in the notch of the cove. The top of member 1 is located on the prominent bench of the promontory. The basal part of member 2 can be viewed along the prominent bench, but the rest of section must be viewed from above (Fig. 2).

Be extremely cautious, the rocks are slippery when wet.

Introduction

The rocks of the Chapel Island Formation on the Burin Peninsula are considered by the International Precambrian-Cambrian Boundary Working Group as the boundary stratotype with additional reference sections in Siberia and southern China. Elsewhere in Newfoundland the boundary is absent or the base of the Cambrian is unconformable. The Chapel Island Formation consists predominantly of relatively fine-grained, storm-dominated shelf deposits in an overall deltaic setting. It rests conformably on the Rencontre Formation, a sequence of red coloured medium, to coarse-grained, cross-bedded quartz arenites, and related siliciclastics, and is in turn conformably overlain by sandstone, shales and quartzites of the Random Formation. In the Burin Peninsula the Chapel Island sequence is about 1000 m thick, and divided into five informal members, numbered 1–5. The Precambrian-Cambrian boundary point is located 2.4 m above the base of member 2.

Description

Upper part of member 1

Three lithofacies comprise member 1: red and green sandstone/shale facies 1.1, shaley facies 1.2, and grey to black shale facies 1.3. Many of the features of facies 1.1/1.2 are readily observed at the base of the section, including synaeresis and desiccation cracks and abundant current ripples. The rocks of facies 1.3 are poorly weathered below the small fault, but contain abundant synaeresis and a few desiccation cracks. The member 1-2 boundary has a number of interesting features including desiccation cracks and abundant phosphatic shale clasts; at the end of the promontory there is a shallow elongate scour that is covered with phosphatic shale clasts up to 15 cm in length. Ripples on the bedding surface adjacent to the scour continue down into the scour, and based on their tuning fork junctions are considered wave ripples.

Facies 1.1 represents a lower tidal flat/shallow subtidal deposit and 1.2 a middle/upper tidal flat deposit. Bedding characteristics, sedimentary structures and paleocurrent data indicate that reversing tidal currents were the main agents of transport and deposition. Facies 1.3 was deposited in semi-restricted shoreline environments of uncertain affinity. Partially enclosed shoreline embayments including interdistributary bay environments are considered probable settings for deposition of this facies. The prominent bench is considered to be a minor disconformity surface that shows evidence of subaerial exposure and reworking by waves after submergence.

Fossils are present sporadically in these beds, mostly in the grey-black units. Trace fossils are simple forms such as *Planolites* and *Gordia*. Well-preserved examples of the carbonaceous small shelly fossil *Sabellidites cambriensis* and the vendotaenid alga *Tyrasotaenia* occur in the darker, shalier units. One impression of a soft-bodied (medusoid) organism was found near the top of these strata. The overall composition of the fauna and flora suggest a latest Precambrian age.

Lowermost member 2 (Precambrian-Cambrian Boundary)

The strata belong to the Gutter Cast Facies 2.1; i.e. mainly laminae to very thin beds of green-grey sandstone and silver-green siltstone, with moderate to abundant gutter casts. The term gutter cast refers to erosional structures that range from downward-bulging sole structures to isolated channels. They are composed of fine sandstone, less commonly of white-weathering well-sorted, fine to medium grained sandstone and granule to pebble conglomerate. The sandstone gutter casts are either massive, planar-laminated, or contain oscillatory-flow laminae. The thicknesses of the gutter casts are generally one or more orders of magnitude greater than the thickness of the sandstone beds with which they are intercalated. Carbonate nodules are found within these beds, and in many cases the entire bed experienced early carbonate cementation. The sandstone content of this facies, excluding gutter casts, is 10–40 per cent. Quartzitic beds make up about 5 per cent of this facies. Sandstone beds greater than a few centimetres are rare. Unifite beds are prominent, forming 10–15 per cent of the section, and one or two raft-bearing beds have been noted. Ptygmatically folded sandstone dikes are surprisingly abundant, in some cases leading to gutter casts. Wave ripples are present on tin sandstone beds, locally as starved forms.

A wide variety of gutter-cast shapes are seen both in cross-section and in plan view. Extremely steep to overhanging walls are common. Compaction has altered the shape and internal stratification in some of these beds. Sole markings on the sides and bases of these beds include groove marks, poorly developed prod and flute marks, and post-depositional trace fossils. Groove marks are generally parallel to subparallel to the long axes of the gutter casts and circular to downward-spiralling on pot casts.

The long axes of gutter casts show a strong northeast-southwest orientation. Some gutter casts are quite sinuous. Numerous examples of bifurcating gutter casts were noted, as well as those that taper and pinch out along strike. All of the bifurcating examples have their forks opening toward the northeast, and nearly all pinchouts are toward the northeast as well. The upper bedding surfaces of some gutter casts are covered with symmetric ripples marks. The crests of these ripples are generally perpendicular to the trend of the gutter cast, but beds with multiple directions occur.

In a few cases, gutter casts are seen originating from, and leading into, pot casts. These pot casts commonly weather out as partially free-standing structures. They range from discs to rounded loaf-like forms to tall pillars and from remarkedly small (1 cm diameter) to very large (ca. 20 cm diameter). They commonly widened downward, and a few examples have a snail-like or corkscrew shape similar to potholes found in bedrock along modern rivers. The bottoms of pot casts are commonly deepest around the outside, with a central erosional high; the form resembles the base of a wine bottle.

Unifite beds are graded to nongraded siltstone and silty mudstone characterized by a lack of obvious internal structure in the field. Cut and polished slabs, however, reveal a range of subtle internal structures. The bases of these beds are very sharp, and the tops vary from gradational to very sharp. These beds range from 10–70 cm thick, averaging 36 cm at this exposure. These beds are normally tabular, but locally have a channelled margin.

Intimate association with peritidal redbeds and the shoreline deposits of Facies 1.3 indicates a shallow subtidal environment. The gravity flows suggest high sedimentation rates and deposition in deltaic environments. High sedimentation rate is supported by the abundance of sedimentary dikes.

Data from this outcrop prompted the model for tempestite deposition along fine-grained shorelines. The gutter cast setting represents a zone of bypass, an area dominated by erosion and throughput of sediment. Perhaps fast-moving sediment flows were debouched from distributary channels into the shallow subtidal environment, where they eroded the sea floor on entry into the marine setting and then deposited their load farther out across the delta-front, prodelta, and inner shelf.

Observations from the gutter casts in this facies, including steep to overhanging side walls, suggest a short period of time between cutting and filling. Overhanging walls must have been very common in potholes because most pot casts are tilted and/or widen downward. These potholes were eroded into semi-cohesive to cohesive substrate of clayey silt by stationary eddies carrying sand in a manner similar to that outlined for bedrock potholes. The thickness of the gutter casts (generally 10-100 times thicker than the associated sandstone beds) and the evidence for rapid infilling of the gutters is compatible with deposition from thick sediment-laden flows that deposited sediment almost exclusively in the gutters. The wave ripples on the top of several gutter casts formed during the late stages of deposition or from later reworking prior to burial.

Unifite beds have grain sizes (dominantly silt) particularly susceptible to liquefaction and downslope flow. The source may have been muddy delta-front mouth bars. The well graded and stratified unifites may have been generated closer to shore, allowing evolution into a turbidity current prior to deposition on the delta front. Excellent analogues for these silty flows have been described from the Huanghe (Yellow River) Delta. They describe silt-flow gullies that are filled with acoustically transparent sediments thought to have been generated by liquefaction.

Fossils are abundant and can be used to define a

Precambrian-Cambrian boundary. Member 1 and the basal 2.4 m of member 2 represent the uppermost strata of the *Harlaniella podolica* Zone. This zone is characterized by a low diversity trace fossil assemblage dominated by simple subhorizontal burrows (*Palnolites, Gordia, Harlaniella,* and *Palaeopascichnus*). The latter two ichnogenera are particularly valuable age indicators as they apparently are restricted to the upper Precambrian. The vendotaenid alga *Tyrasotaenia* is also common in these strata.

Phycodes pedum first appears 2.4 m above the base of member 2, and can be used to define the base of the *Phycodes pedum* zone. This is also the stratotype point and horizon for the Precambrian-Cambrian boundary and occurs within a continuous and relatively uniform succession of subtidal strata. The highest occurrence of Vendian ichnofossils (*Harlaniella podolica* and *Palaeopascichnus delicatus*) is only 0.2 m lower in the section, in the same lithofacies.

Higher strata of member 2 at this location contain elements of the *Phycodes pedum* zone including arthropod traces (*Monomorphichus*), vertical dwelling burrows (*Skelithes* and *Arenicolites*) and coelenterate resting burrows (*Conichnus*). Vendotaenid algae also occur sporadically in these strata.

Source

Myrow, P.M., Narbonne, C.M., and Hiscott, R.N., 1988. Storm-Shelf and Tidal Deposits of the Chapel Island and Random Formations, Burin Peninsula: Facies and trace Fossils. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, St. Johns '88, Field Trip B6.

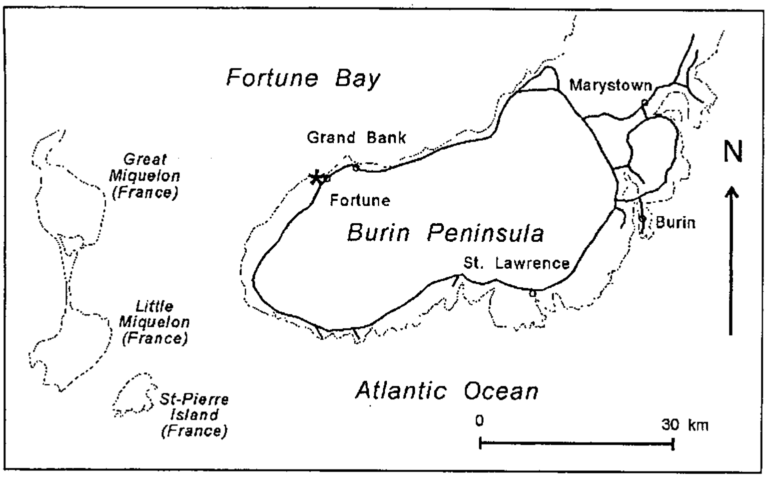

Figure 1: Location map for Fortune Bay.

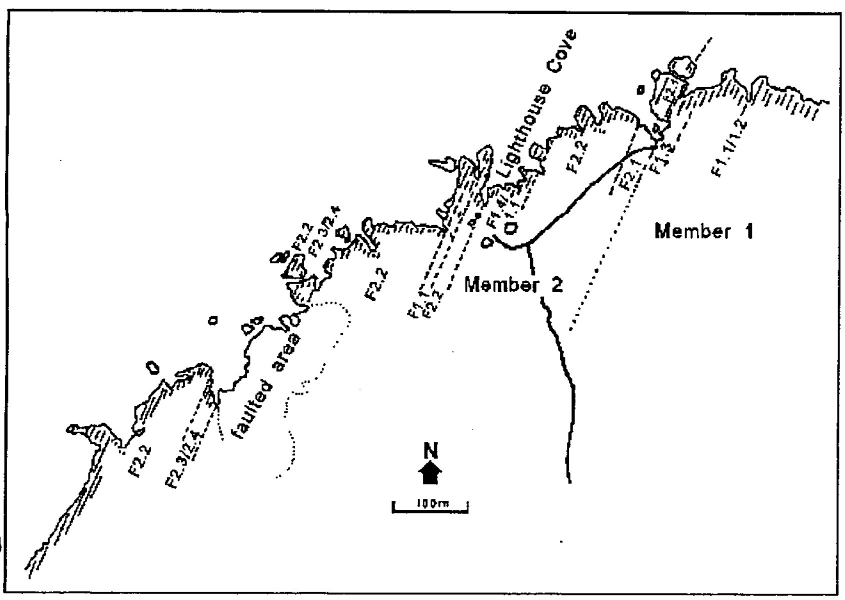

Figure 2: Detailed stratigraphy map of Chapel Island Formation at the Fortune Head lighthouse. The Precambrian-Cambrian boundary is 2.4 m above the base of member 2. The sequences are facies described in the text. From Myrow et al., 1988.

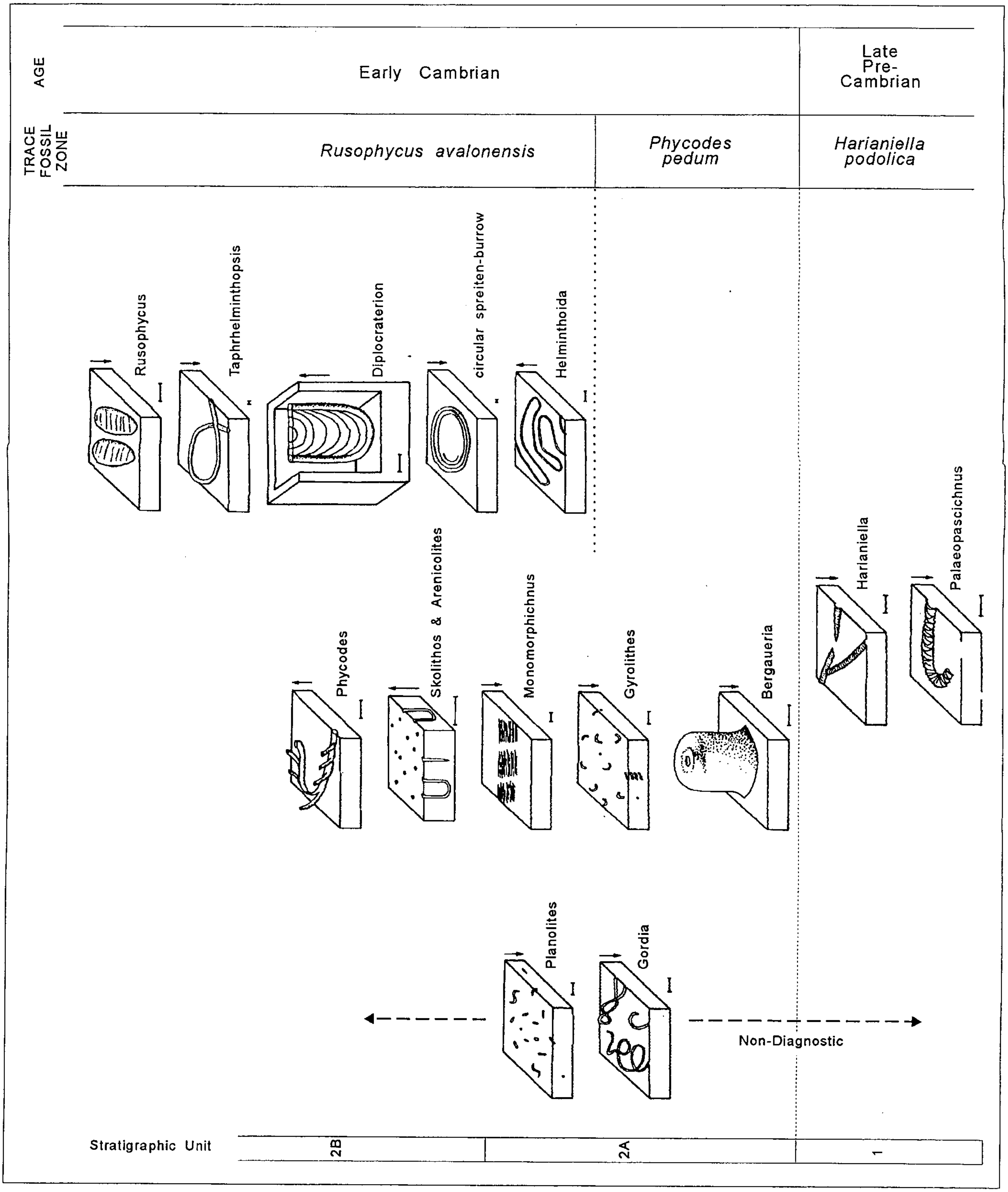

Figure 3: Trace fossil stratigraphy at the Precambrian-Cambrain boundary, Fortune Bay. Modified from Myrow et al., 1988, figs. 13 & 14.

Moreton's Harbour

Pillowed Lava Flows and Volcaniclastic Sediments of the Lower Paleozoic Cover Rocks of the Gander Zone

Purpose

The Little Harbour exposures are characterized by especially striking volcanic features such as the "splash structures" of liquid bombs and intrusions of sills into unconsolidated tuffs, where they contracted into rounded pillow-like forms. The high proportion of epiclastic debris within the Moreton's Harbour Group suggests generation in an island arc environment. The Moreton's Harbour Group lies within the Notre Dame Subzone.

Location

To get to Moreton's Harbour from the TransCanada Highway take either 340 (Lewisporte-New World Island Road) to the village of Virgin Arm or 380 (Gander Bay Road) north to 331 at Victoria Cove and then turn onto 340 at Boyd's Cove to get to Virgin Arm. At Virgin Arm take 345 to Moreton's Harbour (Fig. 1). In Moreton's Harbour turn east just past the museum towards Wild Bight and Tizzards Harbour and then the small road left going around the east side of the harbour. Park near where a picket fence encloses two large satellite dishes and take the trail next to the hillside. The rocks of interest are beautifully exposed around Little Harbour, the small harbour to the northeast.

Description

Moreton's Harbour—Little Harbour

The Morton's Harbour Group is a southwest-facing sequence of pillow lavas, pillow breccias, aquagene tuffs, and minor cherts, which along with the large volumes of diabase dikes comprise a total thickness in excess of 8 km. The group has been somewhat arbitrarily separated (Fig. 1) into formations named from the oldest to youngest as the Tizzards Harbour, Webber Bight, Little Harbour, and Western Head Formations. They are intruded by gabbro and abundant diabase dikes, and the extensive area of diabase dikes separating the Webber Bight and Little Harbour Formations is termed the Wild Bight Intrusive Terrane. Felsite dikes are also common, and they range from early to late in the sequence. The Group, especially the Tizzards Harbour Formation, exhibits a gradational eastward metamorphic transition from greenschist to amphibolite facies rocks in the northeastern part of the area.

The Little Harbour Formation that is exposed along the shore north of Little Harbour has a total accumulated thickness of 2600 m. It consists of close-packed pillow lavas and their feeder dikes, overlain by pillow breccia, coarse-grained aquagene tuff with isolated pillows, and fine-grained and graded tuff, capped by thinly laminated siliceous tuffs. Within the clastic sequence the three varieties of tuff are rapidly repeated but pillow lavas occur only in minor amounts above the main basal lavas. Bedding is characterized either by thin, 1 to 30 cm, fine-grained laminated units or large, 1 to 2 m thick, structureless beds. Clasts up to 20 cm in diameter, irregular to angular shaped, occur throughout; some clasts have bedding draped around them. The dikes and sills are 20 cm to over 1 m thick, have chilled margins, and are tabular shaped. A few have irregular margins. The bedding within the tuffs dip moderately (70°) to the southwest and the dikes dip steeply to vertical, striking to the northeast-eastnortheast.

All known mineralized veins, containing quartz, calcite, arsenopyrite, pyrite, stibnite, pyrrhotite, sphalerite, chalcopyrite, galena, and gold are found within the Moreton's Harbour Group. Up behind the field is an old antimony mine whose waste pile yields good samples of coarsely crystalline pyrite and silvery stibnite. Do not venture into the mine or the trench!

Wild Bight

The Wild Bight Intrusive Terrane consists of an extensive area of more than 95 per cent basic intrusive rocks with only very minor screens of pillow lavas, which occupy a stratigraphic thickness of about 1300 m. The dominant rock type, comprising over 70 per cent of the formation, consists of coarse-grained diabase dikes that strike northeasterly and dip gently (30°) to the southeast. It must be emphasized that, although most abundant in this area, these dikes extend over much of the peninsula, and in some areas bisect the country rocks into slabs—for example, where they cut the vertical sediments of the Tizzards Harbour Formation.

The coarse-grained subhorizontal diabase dikes are cut by fine-grained diabase dikes which are dominantly vertical and strike from northeast to northwest. In the southeast part of Wild Bight, where they strike northeast, these dikes are locally 'sheeted' (i.e. consist of up to 100 per cent dikes) over widths of hundreds of metres. Some dikes are separated by thin screens of pillow lava and pillow breccia which appear similar to those of the Little Harbour Formation.

The coarse-grained subhorizontal sheets are most likely feeders to the volcanic pile, since they become vertical if the bedding is rotated back to an original horizontal position. If this is done the fine-grained sheets are probably feeders to later lavas which followed the tilting of the earlier units.

Source

Williams, H., Steven, R.K., Blackwood, F., Papezik, V.S., and Malpas, J., 1980. A Cross-section Through the Appalachian Orogen in Newfoundland; Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Halifax, '80; Field Trip 14; 63 p.

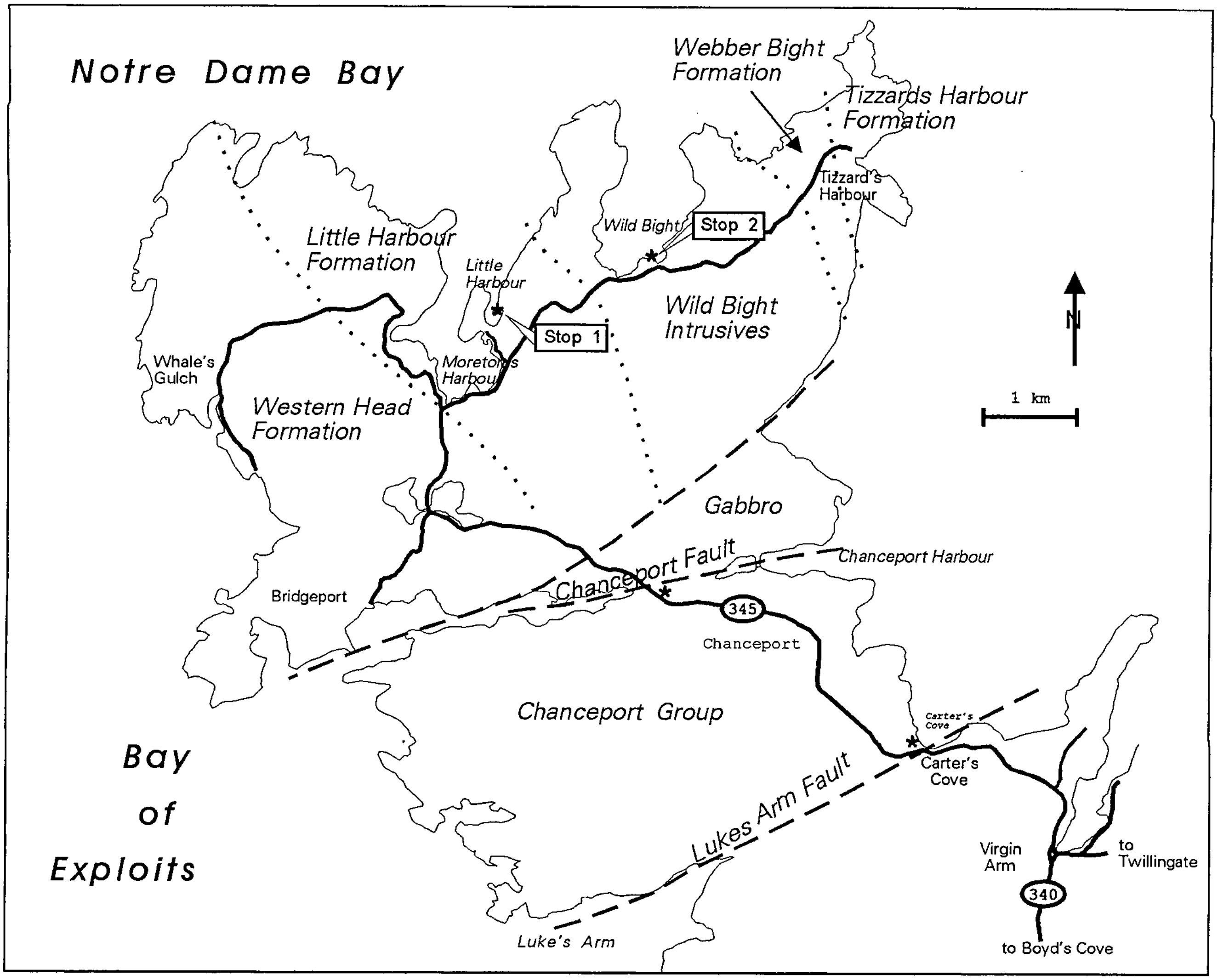

Figure 1: Location may showing regional geology of Moreton's Harbour area. Modified from Williams et al. 1980

St. Anthony

Metamorphism Beneath an Ophiolite

Geological Background

The St. Anthony Complex, located at the northern tip of the Great Northern Peninsula (Fig. 1), is comparable to the Bay of Islands Complex in that it represents a slice of oceanic crust and upper mantle emplaced over the continental margin of Laurentia during the Ordovician Taconian orogeny. However, the St. Anthony Complex differs from the Bay of Islands Complex in that it is older and contains only the lower ultramafic part of the classic ophiolite suite; in addition it contains a very well developed "metamorphic sole" formed during emplacement of the ophiolite. In general terms, ophiolite emplacement is thought to have started when one section of oceanic crust over-rode another along an east-dipping subduction zone. The gabbros, basalts and marine sediments of the downgoing plate were subjected to progressively higher pressures and temperatures, as well as intense deformation. Eventually, a thin slab of highly deformed and metamorphosed oceanic rocks, showing reverse metamorphic zonation from high grade amphibolites at the top to low grade greenschists at the bottom, became welded to the base of the overlying wedge of upper mantle peridotite. This metamorphism has been dated at 490 Ma, about the time of the Cambrian-Ordovician boundary. Later, the upper oceanic plate became detached to form a large ophiolite sheet which was thrust over continental margin sediments, bringing its underlying "metamorphic sole" with it.

The St. Anthony Complex is one of the best preserved examples of a sub-ophiolite metamorphic sole anywhere in the world. It consists of the White Hills Peridotite which is the remains of an ophiolite complex and underlying metagabbros, amphibolites, greenschists, metasediments, and volcanic rocks (Fig. 2). The rocks range from undeformed pillow lavas to polydeformed ultramylonitic rocks. This represents a change in P-T conditions ranging from <350°C, 2 kbar to ca. 900°C, 7-10 kbar pressure, all in a structural thickness of less than 750 m (Jamieson 1981). The St. Anthony Complex essentially represents a shear zone between the crust and the mantle (Jamieson 1986), with a very rapid transition from undeformed to mylonitic rocks, and an increase in metamorphic grade towards the crust-mantle boundary, which is now represented by the contact between the peridotite and underlying pyroxene amphibolites. This field guide describes two important and accessible sections within the St. Anthony Complex—the transition from undeformed pillow lavas to greenschists, and the transition from greenschists to amphibolites.

Loctaion

Take Route 430 (the "Viking Trail") along the west coast of the Great Northern Peninsula to St. Anthony. This is a 5 to 6 hour drive from Deer Lake, or a 4 to 5 hour drive from Gros Morne National Park. About a kilometer past the outskirts of St. Anthony (between the stadium and the shopping mall), a gravel road heads uphill to the west (on the right as you drive into town). This is the road to Goose Cove. Follow this road for about 11 km—there are very few side roads, so you shouldn't get confused. Within sight of the village of Goose Cove there is a fork in the road (Fig. 3)—bear left on the main road to Goose Cove for Stop 1, or take the right fork to Goose Cove West for Stop 2.

Access

Both stops are on public land, so there is no need to ask permission. Access is by foot from the road along poorly defined paths. Make sure your vehicle is pulled well to the side of the road so it does not pose a hazard to local traffic, and to avoid flying stones as much as possible. Potential safety hazards at each stop are described below, but be prepared for rough walking (through bogs or tuckamore) and/or scrambles along steep sections. You are strongly advised to carry a first aid kit in your pack. These stops are not suitable for children, and the descriptions are therefore written for university students and others with some geological field experience.

Stop 1: Three Mountain Summit

Directions

Pull off about 1 km from the fork, or about 500 m uphill from the outskirts of Goose Cove. There is a prominent roadcut on the east side of the road, with a power pole on top of it, and a low-lying alder swamp west of the road. Climb up the side of the roadcut nearest the power pole and follow the rough track towards the highest prominent hill, which is the southwestern "summit" of Three Mountain Summit. Unfortunately, trails that were well marked in the late 1970's were obscure in the early 1990's—local people now use

ATV's to get around, but they don't take them up the hills! You may have to do some bog-walking or bush-whacking to get to the base of the hill. Aim for the northern shoulder of the hill--the rocks are well exposed and there is less tuckamore. The round trip to the top of the hill and back will take 2-3 hours, depending on how long you spend peering at the rocks and admiring the view from the top.

Geology

The St. Anthony Complex is an inverted metamorphic sequence that forms a subhorizontal sheet (see cross-section in Fig. 2). The lowest grade, least deformed rocks are near sea level, and higher grade, strongly deformed rocks form the high ground. The transition from mylonitic greenschists of the Goose Cove Schist to amphibolites of the Green Ridge Amphibolite is exposed on the flanks of Three Mountain Summit (Fig. 3).

The roadcut consists of light to dark green, finely banded, mylonitic Goose Cove Schist. These rocks qualify as mylonite because of their extremely fine grain size, fine-scale, continuous tectonic layering, rare isoclinal folds, and small feldspar porphyroclasts, and the rapid gradations from recognizable metavolcanic rocks into high strain zones which can be seen in some places. The protoliths of these rocks were mafic pillow lavas that can be seen at Stark's Bight (Stop 2).

As you walk uphill you will pass into dark grey, shingly epidote amphibolites, biotite-rich mylonite, and finally medium-grained black amphibolites. In fact, almost all the rocks in this section are mylonites—the epidote amphibolites are the retrograded and further deformed equivalents of the coarser-grained amphibolites, which are in turn mylonitized gabbro or diabase. The biotite-rich mylonite represents a somewhat later phase of deformation accompanied by intense chemical interaction with a potassium-rich fluid (metasomatism). This locally transformed the amphibolites into garnet- and biotite-rich, meta-sedimentary-looking material containing irregular "lumps" of disrupted amphibolite.

At the top of the slope, the hill levels off into a shallow spoon-shaped depression. The rocks at the summit consist of black-and-white, finely banded, protomylonitic amphibolite that lacks epidote. This amphibolite is clearly higher grade than the underlying rocks and represents the protolith of some of the mylonites. The contact between the amphibolite and the underlying mylonitic epidote amphibolites is gently folded into a north-plunging synform running along the top of the hill.

To the south, Three Mountain Summit offers a spectacular view of the village of Goose Cove and across Hare Bay to the Fischot Islands and the Grey Islands. The peninsula beyond Goose Cove consists mainly of greenschists and metasedimentary rocks of the Goose Cove Schist, much lower grade than those you are standing on, and affected by large-amplitude post-metamorphic folds. On the isthmus south of the village is the abandoned Goose Cove copper mine which was active at about the time of the First World War. The ore was never shipped and is still sitting in dumps beside the abandoned shafts. One of the shafts excavated a steeply plunging, tight asymmetric fold, demonstrating that the massive sulphides were deformed with the schists. If you visit the mine site, use extreme caution, as the shafts are not fenced off and are filled with stagnant water!

STOP 2: Stark's Bight

Directions

Take the west branch of the Goose Cove Road towards Goose Cove West (Fig. 3). Stop about 1 km along the road, where it makes its closest approach to the bay on the west (Stark's Bight), and where a small stream winds its way down to the shore. Walk west down the slope to the cove--there may be a poorly defined track, or you may simply have to walk across the ground cover (mainly Labrador tea and caribou moss with a few alders—not too bad). Follow the coast north from the cove. Trails that were easy to follow in the early 1980's were hard to find 10 years later, *so use extreme caution*. This section is not suitable for children, or for those who are nervous of heights. Although there are some very "scrambly" places, for experienced hikers there are no impossibly difficult sections, and ropes or special climbing equipment are not required. An (expensive) alternative is to rent a boat in Goose Cove; get the boatman to take you to the north end of Stark's Bight and to let you out at various places along the shore. The round trip to the end of Stark's Bight and back takes several hours, depending on walking conditions and how long you spend looking at the rocks.

Geology: At Stark's Bight, the transition from undeformed mafic pillow lavas of the Ireland Point Volcanics to the highly deformed Goose Cove Schist is exposed in a beautiful coastal section (Fig. 3). The section is described as you first encounter it—i.e., from south to north, or from schists to volcanics. However, it is actually more impressive walked from north to south, because the obliteration of primary volcanic structures is more obvious when you have seen the undeformed rocks first.

The first rocks you will see are flaky, carbonate-rich schists of the lower Goose Cove Schist. These rocks are very strongly foliated and lack any obvious primary features. As you walk along the coast, you should look for the first evidence that these rocks were originally volcanic, and also look for changes in the types of tectonic structures present. The first clearly recognizable volcanic rock is a pillow breccia that is exposed at the water line about 500 m north along the coast. The pillow fragments are oxidized and contain bright green epidote veins; the original breccia structure is still evident despite strong deformation of the clasts and their matrix.

As you continue north toward the head of Stark's Bight, primary volcanic features, including purple vesicular pillow lavas with abundant inter-pillow carbonate, become more obvious. These rocks are presumably the protoliths of the flaky carbonate-rich shists that you just walked over. At the head of the bay, cross the small stream and walk a short distance to a prominent outcrop of pale green rocks. These are pillow lavas with abundant plagioclase phenocrysts concentrated in the cores of the pillows. The plagioclase was presumably accumulating in a magma chamber immediately prior to eruption. These rocks are virtually undeformed, although they have been pervasively affected by low-grade metamorphism—the pale green colour is largely due to epidote and chlorite that have replaced the original volcanic groundmass. The deformed equivalents of these very distinctive pillow lavas can be recognized in numerous places within the Goose Cove Schist, including at Cremailliere Harbour (see below). Beyond these outcrops, undeformed pillow lava extends for several kilometres along the coast to Ireland Point, where the volcanic section is truncated by the Hare Bay Thrust, the fault along which the St. Anthony Complex was emplaced over the underlying Maiden Point Formation sandstones. Unfortunately, this spectacular structure is accessible only by boat.

Return to the road by retracing your route along the coast—do not attempt to climb the hill, because the tuckamore is extremely thick here. As you walk back along the section, note how the pillow structures are totally obliterated over a structural thickness of only about 30 m.

Other Points of Interest in the St. Anthony Area

Cremailliere Harbour

About 4 km north of Goose Cove and about 10 km southwest of St. Anthony, a small side road branches off to the east from the main Goose Cove road. This leads to Cremaillere Harbour, an uninhabited, well protected bay with what passes for a sandy beach at the end of the road. From the beach a small track leads south up over a small ridge and back down to the water on the south side of Observation Point. The rocks here are the deformed equivalents of the plagioclase-rich pillow lavas exposed at Stark's Bight. They are clearly recognizable, although the pillows are greatly elongated and flattened, and the deformed plagioclase phenocrysts define a prominent stretching lineation.

St. Anthony Lighthouse

At the end of the road on the south side of St. Anthony Harbour, a picturesque lighthouse is built on alkali gabbro that intrudes sandstones of the Maiden Point Formation. These rocks are essentially undeformed and unmetamor-phosed. The gabbro contains phenocrysts of titanaugite and kaerstutite (Ti-rich hornblende) and has petrological affinities with the Ireland Point Volcanics (Jamieson 1977). The hill that rises steeply to the south behind the graveyard consists of strongly foliated and banded greenschists of the Goose Cove Schist. The basal contact between the St. Anthony Complex and the underlying Maiden Point Formation must therefore lie at the foot of the hill, but is obscured by thick vegetation at this locality.

Grenfell Mission

St. Anthony was the headquarters of Sir Wilfred Grenfell, the medical missionary who brought hospitals, co-operatives, and temperance to the coasts of Labrador and northern Newfoundland. Grenfell's house still stands near the large, modern hospital that is named after his successor, Charles Thomas. The craft shop near the hospital sells a range of local products, including Inuit carvings and embroidered parkas made of light-weight, windproof, water-resistant "Grenfell cloth" (the Goretex of its day!).

L'Anse aux Meadows

The only confirmed Norse settlement in North America (not including Greenland) is located at L'Anse aux Meadows, about a 30-minute drive north of St. Anthony.

This was the first official United Nations World Heritage Site, and is well worth a visit. The site, thought to be the "Vinland" of the *Greenlanders Saga*, was discovered and excavated by Norwegian archaeologist Helge Ingstad and his wife Anne Stine in the late 1960's and early 1970's. They had spent many years searching coastal Newfoundland and Labrador for settings matching the descriptions of Vinland in the sagas. The story of their quest is told in the book "Westward to

Vinland" (a far more historically accurate account, although less entertaining, than Farley Mowat's book on the same subject, "West-Viking"). The connection of the L'Anse aux Meadows site with Vinland has been corroborated by the shapes of the building foundations, radiocarbon dates clustering at about 1000 AD, a single Norse cloak pin, and evidence of smelting. A modern interpretation centre and reconstructed Norse buildings are now located at the site.

The turn-off to L'Anse aux Meadows from Route 430 is located about 15 kilometers northwest of St. Anthony. The roadcuts consist mainly of sandstones and greywackes of the Maiden Point Formation, with some outcrops of gabbro and pillow lava. The L'Anse aux Meadows site is located on low-lying boggy ground; the underlying rocks are quite shaley and may be transitional from Maiden Point Formation to broken formation or melange. The bay is very shallow, and would have suited Norse mariners with their shallow-draft vessels.

Although it seems hard to believe from today's landscape, the Norse came to L'Anse aux Meadows for timber. They also smelted "bog iron", making this the first site in North America exploited by Europeans for its mineral resources.

Immediately east of L'Anse aux Meadows, and visible from the site, is Quirpon Island (pronounced "car-poon"). It was presumably named and visited by Basque whalers like those who set up the whaling station at Red Bay on the south coast of Labrador; Jacques Cartier referred to the island by name on his 1535 voyage through the Strait of Belle Isle.

References

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Jamieson, R.A. 1981. Metamorphism during ophiolite emplacement—the petrology of the St. Anthony Complex. Journal of Petrology, 22, 397-441.

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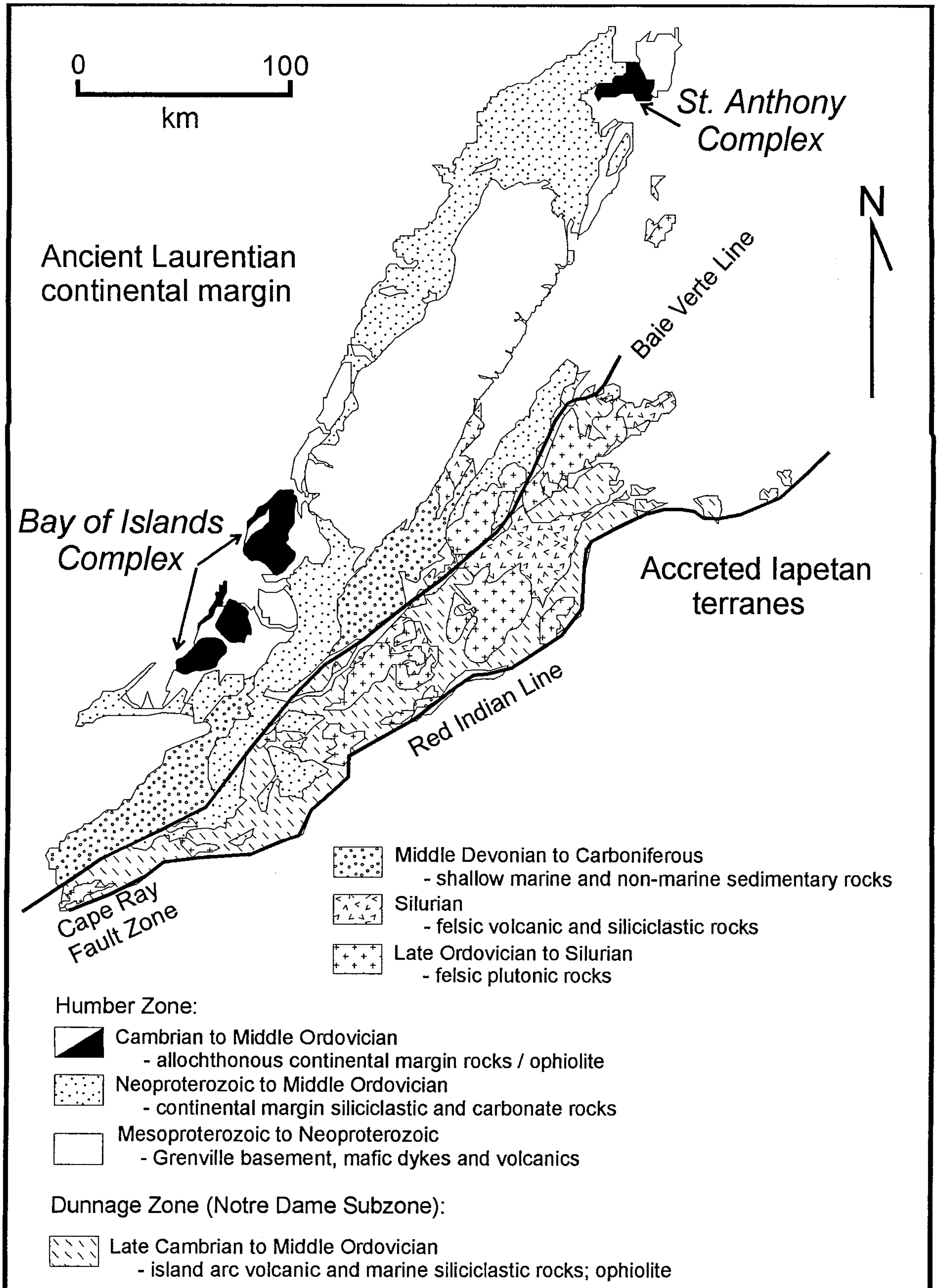

Figure 1: General Geology Map of the Great Northern Peninsula showing the location of the St. Anthony Complex.

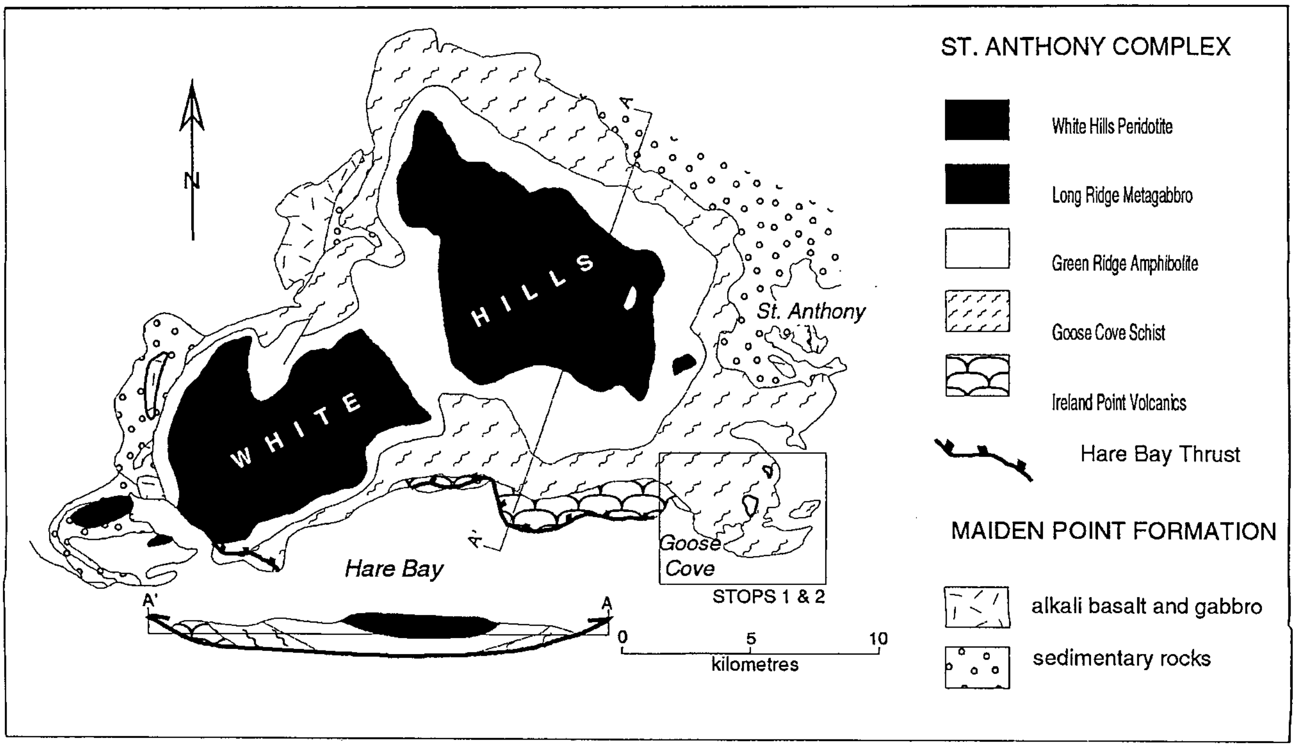

Figure 2: General Geology Map of the St. Anthony Complex

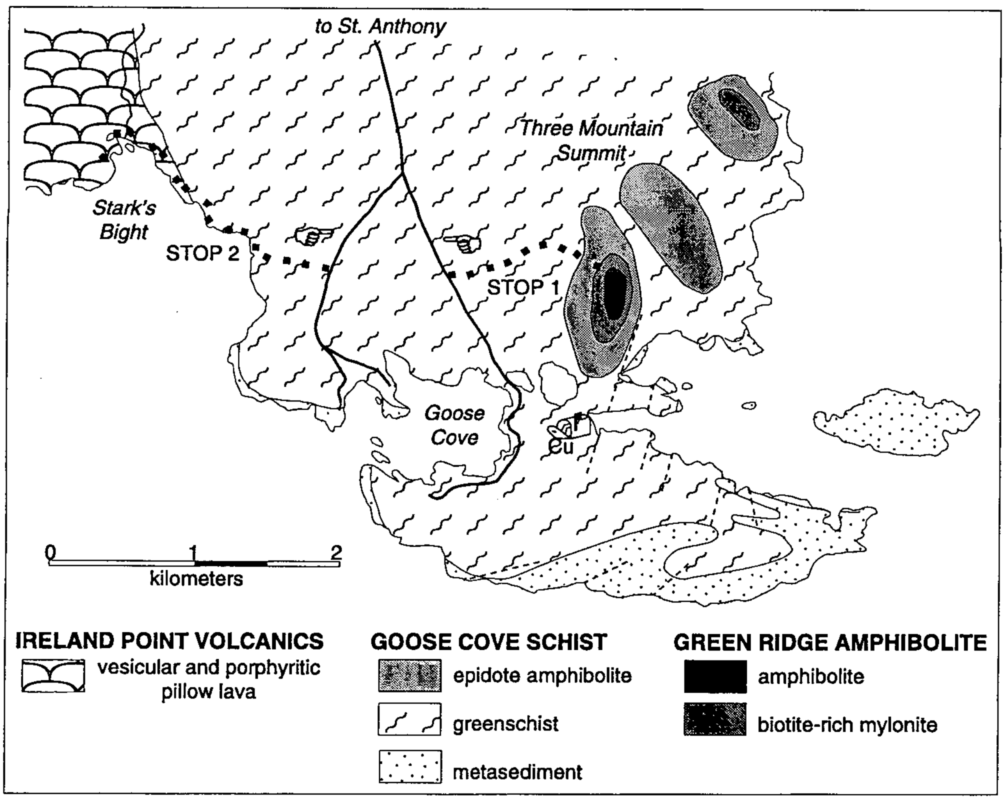

Figure 3: Detailed geology map showing locations of stops in the St Anthony Complex.

Point Riche

Bioherms of the Table Head Group

Purpose

Point Riche contains some of the earliest examples of reef mounds dominated by sponges that appear sporadically throughout the early Paleozoic. These reefs are different from the more familiar Ordovician coral-stromatoporoid bioherms and modern reefs.

Location

Branch off the main road (Route 430) at the town of Port au Choix and drive 4 km to the lighthouse at Point Riche (Fig. 1). This location is part of a national park, its importance derived from the archeological evidence of Maritime Archaic and Paleoeskimo peoples. In light of this, plan a stop at the museum along the road to Point Riche. The point is exposed with little vegetation or shelter, and the waves are energetic and unpredictable. This makes the rocks dangerous so caution must be exercised at all times. Dress warmly and do not go near the shoreline

Introduction

The sequence of rocks at Port au Choix peninsula consists of the upper part of the Lower Ordovician St. George Group and basal part of the Middle Ordovician Table Head Group (Table Point Formation). The Table Head Group is important because elsewhere on the North American craton the Lower to Middle Ordovician is characterised by a hiatus and here this unconformity is either minimal or nonexistent. In western Newfoundland this period is represented by a thick sequence of carbonate rock, in places nearly 2 km thick. The oldest formation in this sequence, the Table Point Formation which is 295 m thick at this locality, contains a distinct horizon characterised by sponge bioherms or reefs.

The Ordovician is an important period in the evolution of reefs. Mounds similar to modern reefs and composed of a diverse group of calcium carbonate skeletons including tabular, cylindrical, and dome shaped colonies first occur in the Middle Ordovician although the basic architecture of reefs was established in the Lower Cambrian. Before the Middle Ordovician, however, most were generally small mound-shaped bioherms or mud dominated highs off the sea floor.

The rocks here lie at the western limit of Appalachian deformation and so are virtually flat-lying, cut only by high-angle faults. The faunas in the Table Head are characteristic of the Whiterock or Llanvirn Stages of the earliest Middle Ordovician in North America or Europe respectively.

Description

On the shoreline near Point Riche (Fig. 1) are exposed some of the best examples of small sponge bioherms. These structures are almost indistinguishable from the surrounding sombre grey limestones of the Table Point, which is probably the reason for them going unnoticed for so long. Believe it or not they are best viewed when wet, that is, in the rain. They are recognized by the presence of fossils (most easily recognized are the gastropods), their lighter colour, and slight deviation in bedding character and orientation.

Fig. 2 is a generalized diagram showing the typical facies of the majority of bioherms. The limestone beneath the bioherms are mostly grey, hackly-weathering bioclastic wackestones, rich in typical shallow-water open-marine fossils. Intercalated in these beds are lenticular skeletal grainstones with the same fauna. These lenses have a concave lower surface and a planar or rippled upper surface. The ripples are symmetrical, with wavelengths of up to 50 cm, amplitudes of 10 cm or less, and crests oriented northwest-southeast. The lenses, which may be up to 5 m in length but rarely more than 20 cm thick, are most likely channel fills in an otherwise low-energy, shallow-water wetting. These sediments are underlain by a 40–50 cm-thick argilliferous lime mudstone.

The bioherms are initiated on bioclastic lime grainstones. These sediments are in the form of sand sheets 10–30 cm thick in some areas while in others they are localized accumulations directly beneath the bioherm. The upper surfaces of these grainstones, where visible, are seen to be rippled, with the troughs of the ripples sculptured with burrows and trails.

The bioherms are 0.2–1.0 m high, 0.5–6 m in diameter, circular to oval in plan, and hemispherical (convex-upwards). They appear as irregular areas of massive, light grey lime mudstone encased in darker grey, thin-bedded, muddy limestone. At first glance they are almost indistinguishable from the monotonous succession of dark grey fossiliferous lime wackestone and packstone that make up the surrounding rock.

The bioherms and surrounding wackestones and lime mudstones are capped by a thin (up to 10 cm thick) burrowed lime mudstone with an intraclastic grainstone veneer. This conspicuous rock is in turn covered by thinly bedded, burrowed lime mudstone to wackestone and intercalated yellow-orange weathering dolomite.

The well-exposed small mounds are mostly skeletal floatstones with skeletal boundstones common in their upper parts. Skeletal remains constitute 40–70 per cent of any individual mound and consist mainly of lithistid sponges (*archaescyphiidae*, *aulocopiidae*, *anthaspidellidae*, *eospongiidae*) together with scattered orthid brachiopods, nileiid and illaenid trilobites, trepostome bryozoans, planispiral gastropods, and rare orthocone cephalopods. Skeletal algae are not obvious in outcrop but are commonly found in thin sections of the fine-grained matrix. These algae, together with pelmatozoan debris, are abundant both within the mound matrix and in the basal bioclastic grainstone sheets that underlie and flank the mounds. The upper surfaces of many Table Point bioherms are brecciated, and grainstone-filled channels are found between many of the structures.

From base to top of mounds, four growth stages can be recognized (Fig. 3); (1) pioneer stage—rippled pelmatozoan-rich sands provide foundation for mound growth; (2) colonization stage—single stick and branching archaeoscyphid sponges initiate mound growth; (3) diversification stage—club-shaped, funnel-shaped, domal and massive anthaspidellid, archaeoscyphiid, eospongiid and aulocopiid sponges together with orthid brachiopods and branching to encrusting laminate trepostome bryozoans form the bulk of individual mounds, and (4) domination stage—plate-shaped to cup-shaped anthaspidellid and aulocopiid sponges encrust high-relief areas.

A thin, horizontal, burrowed lime mudstone with an intraclast grainstone veneer overlies the bioherm horizon at Point Riche. The horizontally oriented burrows in this capping lime mudstone have sparry calcite or coarse-grained dolomite cores. These cement-filled cores indicate that the burrows remained open for some time and that surrounding lime mud must have been sufficiently coherent to prevent collapse of the burrow form. Overlying this burrowed horizon is a 12-20 m sequence of thinly bedded, burrowed (again with cement-filled cores) lime mudstone with intercalated layers of yellow-weathering dolomite. Some of the dolomite layers have internal, hummocky, and planar millimetre-thick laminae. Some of the lime mudstone beds show cracking.

The brecciated tops may have formed through reworking of biohermal material and surrounding beds in the zone of wave action and the grainstone channels may be tidal-surge channels. The burrowed lime-mudstone capping the bioherm horizon may have been alternately wetted and dried causing partial lithification during periodic subaerial exposure and can account for the fact that the burrows remained open to later cement-filling. Likewise some cracks in the overlying lime-mudstone beds may be desiccation features. The hummocky dolomitic laminae are identical to stromatolitic layers of tidal-flat facies also indicating shallow water to subaerial exposure. These features, taken together with the vertical zonation in sponge morphology as outlined in the pioneer to domination stages of biohermal growth, suggest strongly that building of bioherms above the surrounding sea floor led to shallowing conditions and intermittent subaerial exposure. Ultimately the shoaling conditions may have resulted in the termination of sponge bioherm development. Following this, the depressions between the bioherms were infilled by sediment and then tidal-surge channels cut into the unconsolidated or weakly lithified inter-mound sediments, as well as into the bioherms themselves.

Source

Klappa, C.F. & James, N.P. 1980. Small lithistid sponge bioherms, early Middle Ordovician Table Head Group, western Newfoundland. Bull. Can. Soc. Petrol. Geol., 28, 425-451.

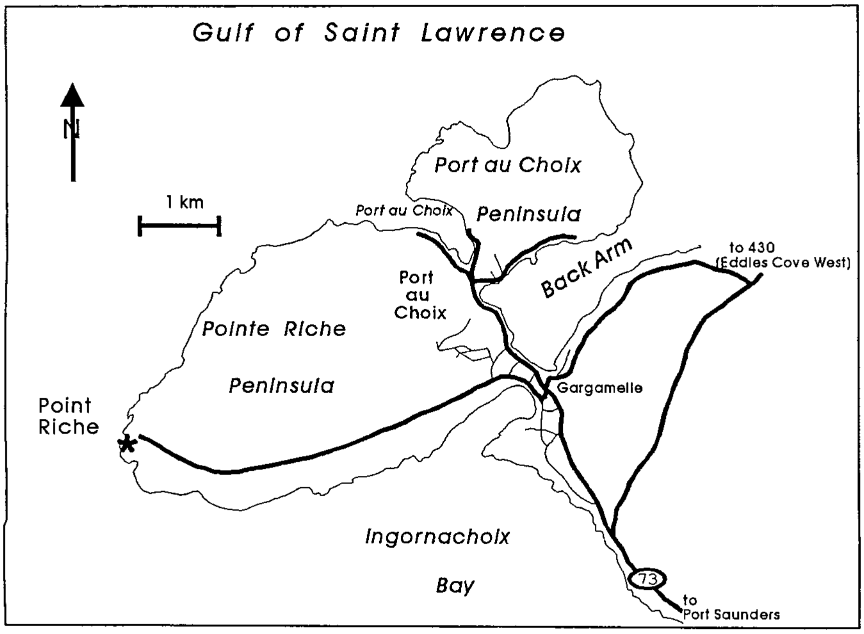

Figure 1: Location map of Point Riche, Newfoundland.

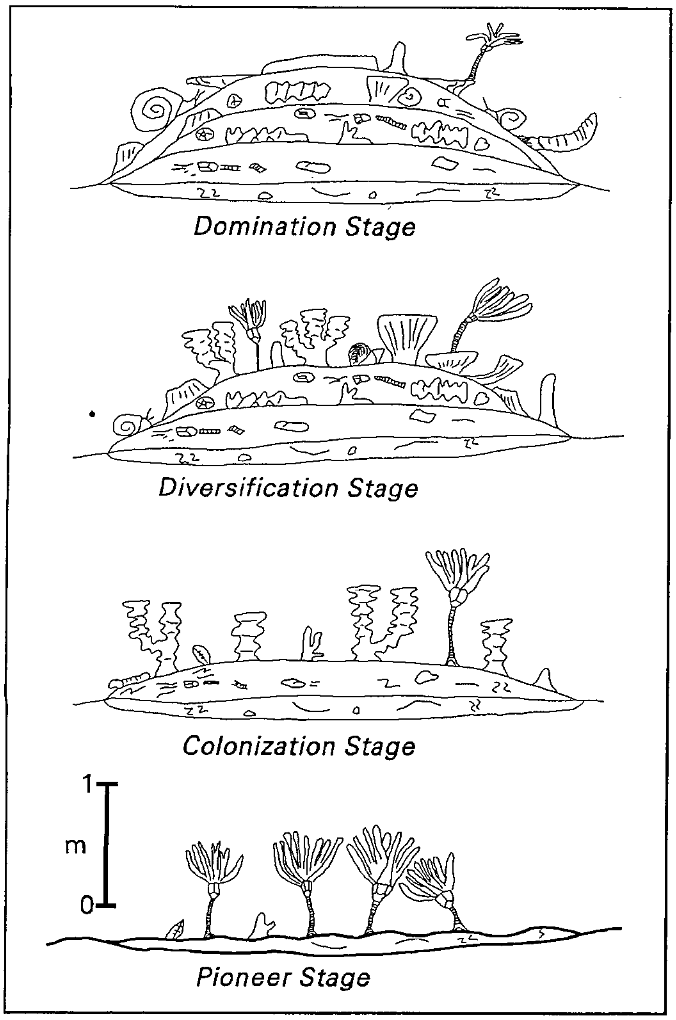

Figure 3: Generalized diagrams of the four stages of bioherm growth at Point Riche. From Klappa and James 1980.

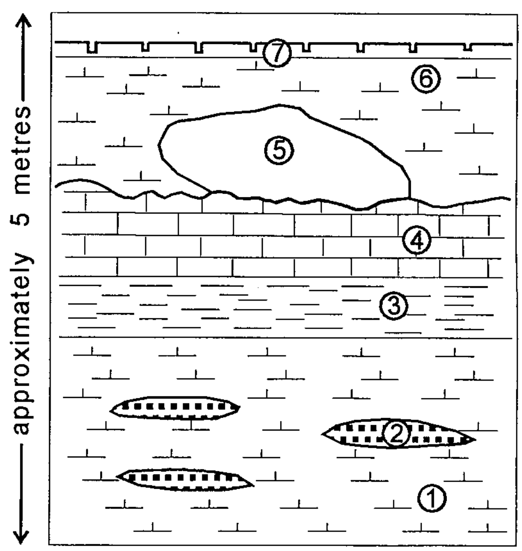

Figure 2: Generalized diagram showing the typical facies underlying, adjacent to, and overlying the majority of the bioherms at Point Riche. 1 - skeletal wackestone; 2 - intraclast bioclastic grainstone lenses; 3 - recessive shaly lime mudstone; 4 - rippled bioclastic grainstone; 5 - bioherm of sponge-algal floatstone to boundstone; 6 - bioturbated skeletal wackestone to lime mudstone; 7 - burrowed lime mudstone with intraclast with grainstone veneer. From Kappla and James 1988.

Cow Head

Limestone Boulder Conglomerate

Purpose

The peninsula of Cow Head is composed of spectacular conglomerates made up of a chaotic collection of limestone fragments, pebbles, and blocks formed partly from underwater landslides.

Location

Exit from highway 430 and drive across the causeway to the peninsula of Cow Head (Fig. 1). Park by the wharf in Tucker's Cove and walk to the north shore of Beachy Cove. An old road running west from the wharf to the lighthouse, where the rocks are Beds 5 and 6, offers another good place to begin.

The best exposures are along the beach and in the cliffs. The cliffs are extremely dangerous as are the tides. Keep well back from the cliffs. View them from afar. There are very few places allowing access to and from the beaches so know the state of the tide at all times and exercise caution when venturing out to view these rocks.

Description

The rocks at Cow Head are composed of the Cow Head Group, part of an allochthonous terrane originally deposited in deep water, as a continental slope to the rise prism of the Iapetus Ocean and transported westward to its present position during the Taconic Orogeny in Middle Ordovician time. The sequence is now a succession of imbricate thrust slices, each of which comprises a lower unit of Cow Head Group conformably overlain by an upper unit of flysch called the Lower Head Formation. The basal contact of the Cow Head and upper contact of the Lower Head are always tectonic. Conglomerates of the Cow Head Group can be classified into five facies (A to E), based upon grading, sedimentary structures, matrix content, sorting, fabric, and clast type (Fig. 2). Facies A consists of graded-stratified, grainy, cobble to pebble conglomerates that have cross-stratified and ripple-stratified tops. The layers are a maximum of about 1 m thick. Sand-sized grains are dominantly peloids, with minor quartz, ooids, and bioclasts.

Facies B, C, D, and E are debris flows. The bases of some facies D and E layers are erosive but most facies B, C, D, and E conglomerates have flat bases showing no more than a few centimetres of downcutting into underlying beds, primarily by plucking of the underlying substrate. Recent studies suggest these facies are simple toe-of-the-slope deposits which accumulated on a southeastward-dipping slope. The Cow Head Group ranges in age from late Middle Cambrian (*Bathyuriscus-Elrathina* zone) to late Canadian, early Whitrockian (Arenig).

The section on Cow Head Peninsula is virtually continuous and represents some 70 million years of deposition. The following description groups the rocks into 5 areas for the sake of simplicity (Fig. 3) starting with the oldest beds at Beachy Cove. The bed numbers refer to groups or packages of individual beds with similar lithologies, sedimentary structures, matrix content and fossils, and were established early in the work on these classic proximal debris low deposits.

Beachy Cove: Beds 1 - 5 (Figs. 1 & 3)

The oldest beds at Cow Head form the cliffs and low tide platform at Beachy Cove. The oldest conglomerates (late Middle Cambrian) are exposed on White Rock Islets to the north in Shallow Bay and the strata at Beachy Cove are earliest Late Cambrian in age.

Bed 1: Several welded conglomerates, composed of small equant to tabular clasts. The most conspicuous boulders are tabular slabs of fossiliferous grainstone.

Bed 2: A relatively thin, resistant conglomerate bed with similar small clasts but exhibiting a calcite spar cement which is locally isopachous.

Bed 3: Present both as a recessive channel in the low-tide zone and as the cliff on the south side of the cove. This is a spectacular deposit consisting of a lower green shale overlain and partly cut out by a thick megaconglomerate. The chaotic megaconglomerate has a shale matrix and is composed of huge coherent but folded masses of parted slope limestone that break up laterally and grade into boulder conglomerate. The unit can be traced from almost 0.5 km along the strike and is capped sporadically by graded calcarenite.

Bed 4: Two gravity flows separated by a thin parted mudstone to packstone. Conglomerates are composed of relatively small clasts but exhibit occasional large boulders and have a buff-weathering dolomitic matrix.

Bed 5: A series of similar small clast conglomerates which are often welded and sometimes separated by parted mudstone and shale. The top of this unit is here limited by a fault and part of the section is repeated to the west.

Lighthouse Flats: Bed 6 (erosional platform opposite the lighthouse)
Bed 6: The low platform formed by Bed 6, although broken and repeated by faulting, is a continuous section some 70 m thick. It is characterized by a wide variety of lithologies but few massive breccias. Strata are distinguished by abundant quartz sand which reaches a maximum at the center of the bed as sandstone layers.

The most common lithologies are quartzose calcarenites with all the characteristics of carbonate turbidites (Facies A).

A peculiarity of the thicker small-clast conglomerates of calcarenites is large isolated boulders "floating"in the bed.

The upper half of Bed 6, which is cliff-forming, exhibits more thick conglomerates, fewer graded calcarenites and more parted to ribbon mudstones.

The age of Bed 6 ranges from medial Late Cambrian (*Aphelaspis-Dunderbergia* zone) to latest late Cambrian (*Saukia* zone).

Point of Head: Beds 7-9.
Bed 7: Upward in Bed 6 there is a gradual increase in the number of conglomerates containing large boulders of white limestone. These culminate in Bed 7, a cliff-forming mass some 16 m thick, comprising 3 welded conglomerates characterized by conspicuous, large, white limestone boulders. The unit cuts down some 4 m into underlying beds. The white boulders, which occur throughout the Cow Head but are largest here, are composed of *Epiphyton* and *Girvanella* together with internal sediment and cement. These blocks are interpreted as fragments of a long lasting upper slope shelf margin facies.

Bed 8: The actual Point of Head is formed by Bed 8 which is basal Ordovician in age. This unit is transitional, the lower part composed of graded calcarenites similar to Bed 6, yet punctuated by coarse conglomerates as is Bed 7, while the upper part exhibits many features that characterize subsequent strata.

In this bed, for the first time, chert occurs both in beds and as clasts; distinct trace fossils are found in abundance; phosphate (collophane) granules are dispersed throughout many conglomerates and brightly hued shales separate carbonate beds. Shallow-water clasts in this unit appear to have undergone more intensive or extensive diagenesis compared to those in older units.

Bed 9: The cove east of the head is formed by an impressive section of parted to ribbon limestones which are medial Early Ordovician in age, equivalent to the Catoche Formation of the St. George Group on the platform. The very evenly-bedded, parted to ribbon lime mudstones to wackestones are separated by black to green shale. Within the limestones are abundant sponge spicules, together with occasional inarticulate brachiopods, hexactinellid sponges and graptolites. Some beds are intensively burrowed and illustrate numerous U-shaped *Arenicolites* while a few show *Zoophycus*.

The Ledge: Beds 10-14
Bed 10: This impressive mass flow deposit exhibits great variation in thickness along strike, from 0.8–7.2 m over 50 m laterally. Clasts comprise all manner of limestones with large ones up to 4 m across. The matrix is argillaceous and the fabric is swirled and chaotic. The top of the bed is capped by a grainstone. The uppermost surface displays two aspects also found in the subsequent conglomerates of Beds 12 and 14, a brown-weathering chert cap and large boulders whose tops have been planed off flat—either by erosion, or corrosion, or what. There is a Dorset Eskimo site on the grassy terrace above Bed 10.

Bed 11: This unit, like Bed 9, is mainly fine-grained sediments. The basal part is a distinctive condensed sequence of thin-bedded, dark red to black, siliceous shale, brown-weathering chert, and lenticular phosphate conglomerate. The upper half of the bed are a series of parted to ribbon limestones. Punctuating the sequence are beds of buff-weathering burrowed dolomite. Sponge spicules and graptolite remains are common. The graptolite faunas from this bed indicate a latest Early Ordovician age.

Bed 12: This massive megabreccia with clasts up to 20 m in size is similar to bed 10 in the chaotic nature of the fabric and wide variety of clasts. The matrix is, however, green shale. Clasts illustrate extensive diagenesis and contain a diverse fauna of trilobites, brachiopods, gastropods, and sponges. Trilobites and brachiopods from boulders in this bed indicate an earliest Middle Ordovician age.

Bed 13: The ribbon to parted limestones that characterise the lower part of bed 13 are peliodal to intraclastic grainstones, which are commonly spiculitic and have silicified tops. Synsedimentary dikes and other injection features are found in some localities.

Bed 14: This uppermost megabreccia is much like Bed 12 but thicker (15 m) and cuts down from rounded shallow-water boulders to twisted masses of slope carbonates. The green argillaceous matrix appears to be at least in part derived from green-grey argillaceous limestones that were soft when redeposited.

Brachiopods and trilobites from boulders in Bed 14 indicate a basal Whiterock (early Middle Ordovician), *Orthidielle* zone, fauna.

Factory Cove is formed by a syncline with Bed 14 and underlying strata occurring again on the south shore. The uppermost strata here are greywackes of the Lower Head Formation exposed in the cove only at lowest tide.

Beck Cove to Jim cove: Beds 9-14

These beds are correlated to those already described at Point of Head and the Ledge (Fig. 3), their repetition due to the syncline at Factory Cove. Bed 10 is the most prominent traceable bed along the cliff and varies in both thickeness and lithology.

Source

James, N.P., Knight, I., Stevens, R.K., and Barnes, C.R. 1988. Sedimentology and Paleontology of an Early Paleozoic Continental Margin, Western Newfoundland. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, St. John's '88, Field Trip B1.

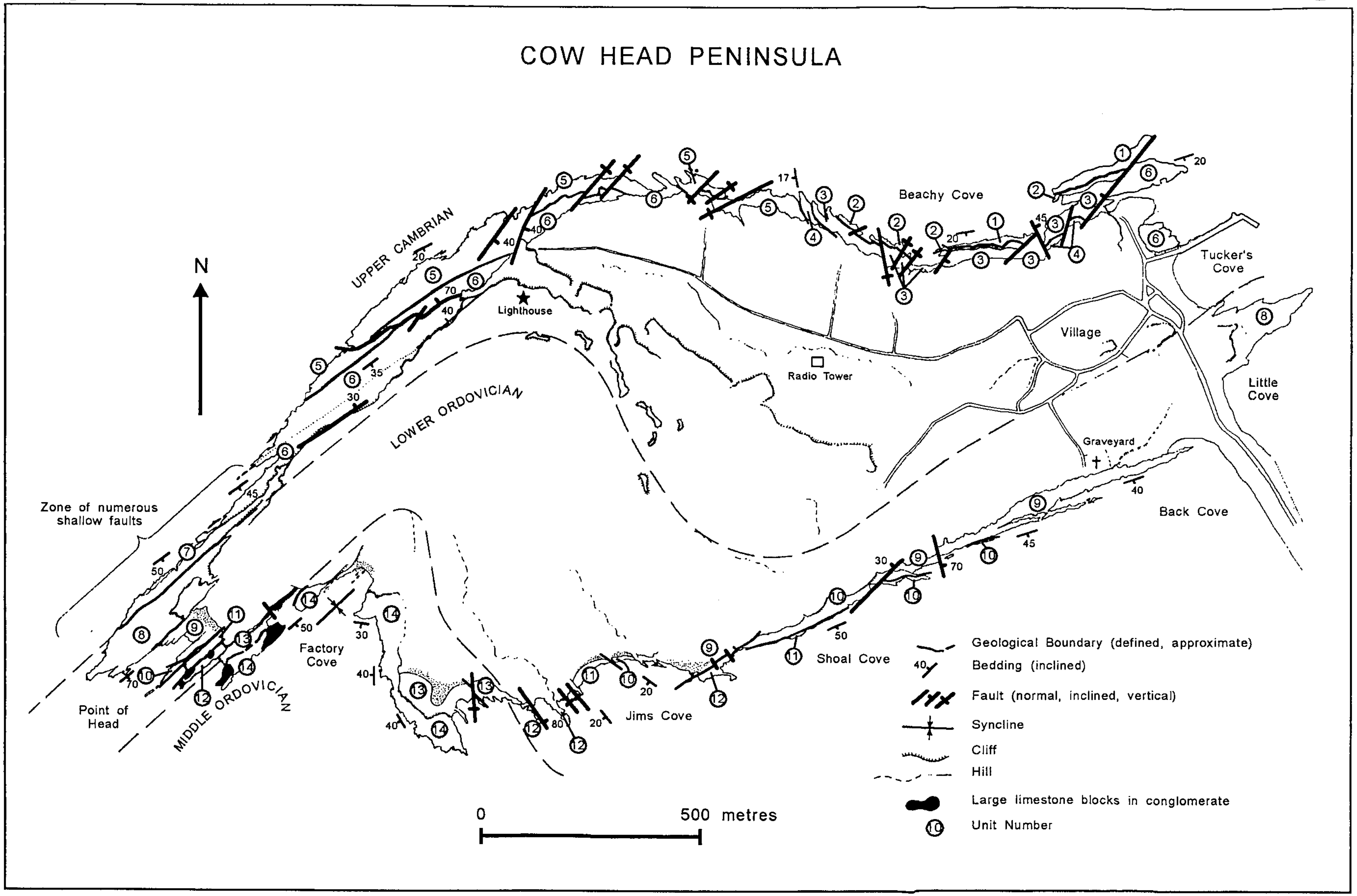

Figure 1: Location map showing local geology of the Cow Head Peninsula. Numbers refer to beds described in the text. From James et al., 1988.

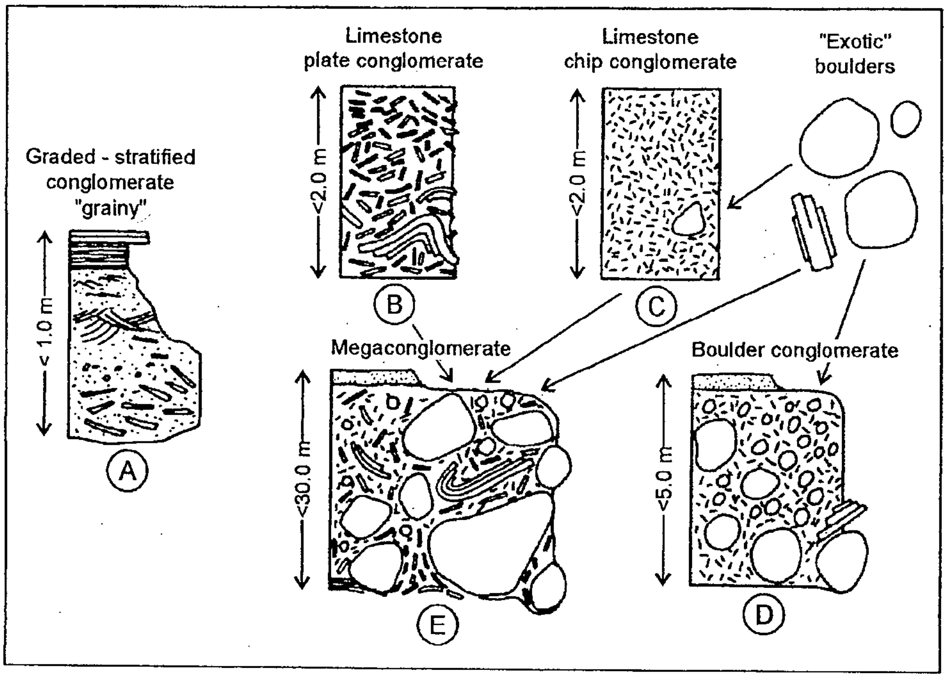

Figure 2: Classifiation of conglomerate facies in the Cow Head Group

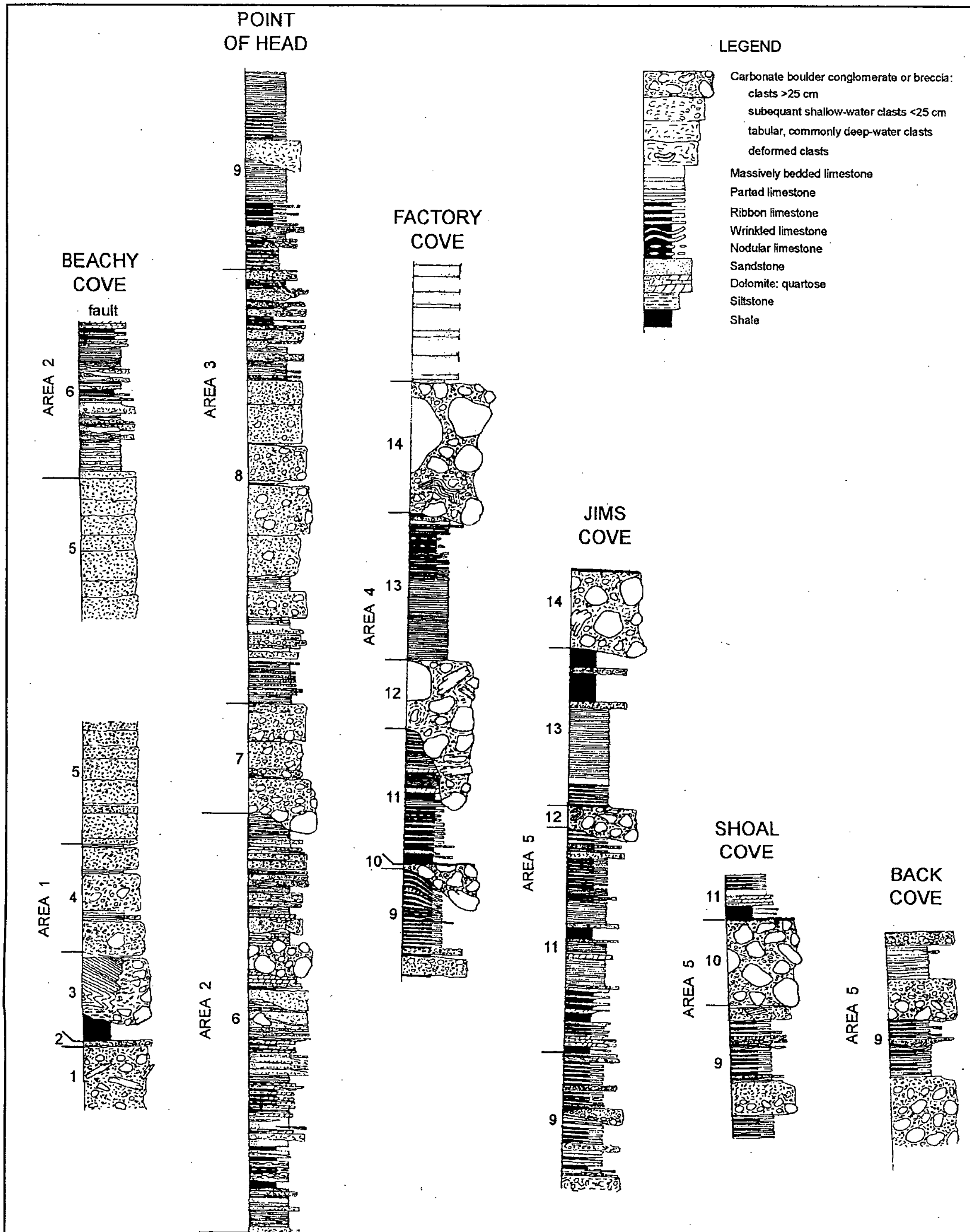

Figure 3: Stratigraphic sections of the Cow Head Group at Cow Head. Note the variable lithologies throughout as well as the downcutting nature of the boulder conglomerates.

Indian Head

Anorthosite and Gneiss of the Grenvillian Basement

Directions

From the TransCanada Highway take the exit towards Stephenville at Route 490 and go to Stephenville Crossing. If you are going south on number 1, take the 460 to Whites Pond and then 461 to Stephenville Crossing. Go west on 490 from Stephenville Crossing (Fig. 1), past Rothesay Bay. About 3 km past the 461-490 intersection as you are going down the hill towards the newsprint plant (past the last but one outcrop on the north) there is a small dirt road to the north. Take this road and park in the small quarry. Walk back to 490.

The other site worth visiting is by the plant. Take the exit to the paper mill and then take the first left fork towards the communication tower to the southeast outside the mill. Drive down as far as you can. Walk down the road to the first small quarry.

If you have the time and inclination the coastal exposures at Indian Head proper are worth the visit. Make sure the tides are low.

Introduction

Anorthosite and gneiss typical of the Grenvillian crystalline basement is exposed at Indian Head and it was upon these rocks that the continental margin was developed during the pre-tectonic Cambrian. Anorthosites are rocks in which the dominant mineral phase is plagioclase. "Plutonic" anorthosites (vs. those of layered basic complexes such as the Stillwater type) are not well understood and their origin is one of the major problems of silicate petrology. They are found on nearly all shield terrains but are most pronounced in the Grenville structural province of Canada where approximately one-fifth of the total surface area is anorthosite—with single massifs covering up to 20 000 km^2. Economically, they are very important for their Ti-Fe deposits (see Fig. 1 for location of mine site).

There are two major types of plutonic anorthosites—the Adirondack type, dominated by very coarse grained andesine which is commonly deformed, and associated with gabbroic anorthosite, norite, and charnockitic (pyroxene-bearing) granitic gneisses; and the gabbroic or troctolitic type, dominated by labradorite and commonly showing cumulate structures including mineral stratification. Arguments concerning anorthosite origin generally focus on whether the rocks represent late liquids from highly differentiated gabbroic intrusions, remobilized plagioclase-rich cumulates intruded as "crystal mush", the residue or partial fusion of quartz diorite with charnockites representing the melt fraction, or an immiscible liquid derived by fractionation of a gabbroic parent melt.

The Indian Head Anorthosite Complex

The rocks of the Indian Head Range are anorthosite, located at Indian Head, and gneisses of the granulite and almandine-amphibolite facies; the gneisses may be divided into those which contain quartz and feldspar as the essential minerals (granitic gneiss), and those in which only plagioclase and pyroxene are essential (basic gneiss) (Fig. 1). Several Fe-Ti ore deposits in the gneisses have been mined from time to time south of Long Gull Pond, the most recent during World War II. The complex protrudes through a cover of recent sediment; a narrow outcrop of the Carboniferous Codroy Group overlies the gneisses around the edge of Rothesay Bay. K-Ar ages of 830 Ma for biotite from gneiss and 900 Ma for biotite from a discordant pegmatite vein suggest that the complex cooled at the end of the Grenvillian orogeny.

Indian Head itself is composed of anorthosite. The greater part (80 per cent or more) of the rock consists of coarse-grained plagioclase crystals up to 15 cm long. These are accompanied by very large interstitial crystals of orthopyroxene, which have crystallized from the residual liquid to produce an ophitic texture; hornblende, magnetite, and ilmenite occur as less important accessories. The rock may locally show coarse flow banding—bands of relatively pure plagioclase alternate with bands rich in orthopyroxene. Dikes and veins of anorthosite up to 20 cm (6 inches) are found cutting the anorthosite mass itself and basic gneiss in the Indian Head and Upper Drill Brook Mines.

The basic gneiss is thought to be part of the noritic margin of the main intrusion. It outcrops right around the margin of the anorthosite mass except for a short distance where the latter is in fault contact with granitic gneiss. It contains bands of magnetite and ilmenite and is intruded by numerous pegmatite dikes. The southern part of Gull Pond is bounded by an outcrop of basic gneiss which is cut off from the Indian Head Mine by a band of granitic gneiss. Between Gull and Oxback Ponds much of the basic gneiss contains potassium feldspar instead of (or as well as) plagioclase, and quartz may be present.

Granite pegmatites are found cutting all the rocks of the area except the anorthosite. Generally they occur as veins about 7-15 cm in thickness but they may impregnate the rock to such an extent as to form dikes up to a metre or more wide. The major constituents of the pegmatites are quartz, perthite, and plagioclase, with biotite and iron oxides as important accessories.

The rocks of the area contain the mineral assemblages of the granulite and probably of the almandine-amphibolite facies. The distinction between the two facies is considered to be due to variations in the amount of water present in the rock during metamorphism; these are variations of bulk composition and not of a physical variable. Therefore distinction between the facies in this area is meaningless.

Highway 490 outcrop

This outcrop is mostly fine- to medium-grained basic gneiss with pods, and dikes of pink granitic pegmatite, light granitic dikes and foliated felsic gneiss. Pods and stringers of coarsely massive- to fine-grained magnetite up to 30 cm thick and several metres long occur throughout and parallel to the gneissic layering. Most of the basic gneiss is foliated, dipping shallowly to the northwest; the felsic rocks show variable foliation. The outcrop is cut by steep north-northeast to south-southwest fractures which increase in intensity to the east abruptly terminating the magnetite pods. The outcrop to the east is mostly granite gneiss and pegmatite dikes.

The Quarries

The outcrops and boulders in the two small quarries, which were formerly mines, are interesting because they contain magnetite and very coarse-grained anorthosite in host basic gneiss. The plagioclase crystals in the anorthosite are up to 3 cm long, the magnetite coarse- to fine-grained.

Source

James, N.P. and Stevens, R.K. 1982. Anatomy and evolution of a lower Paleozoic continental margin, western Newfoundland. Guide book for I.A.S. Excursion 2B, August-September 1982.

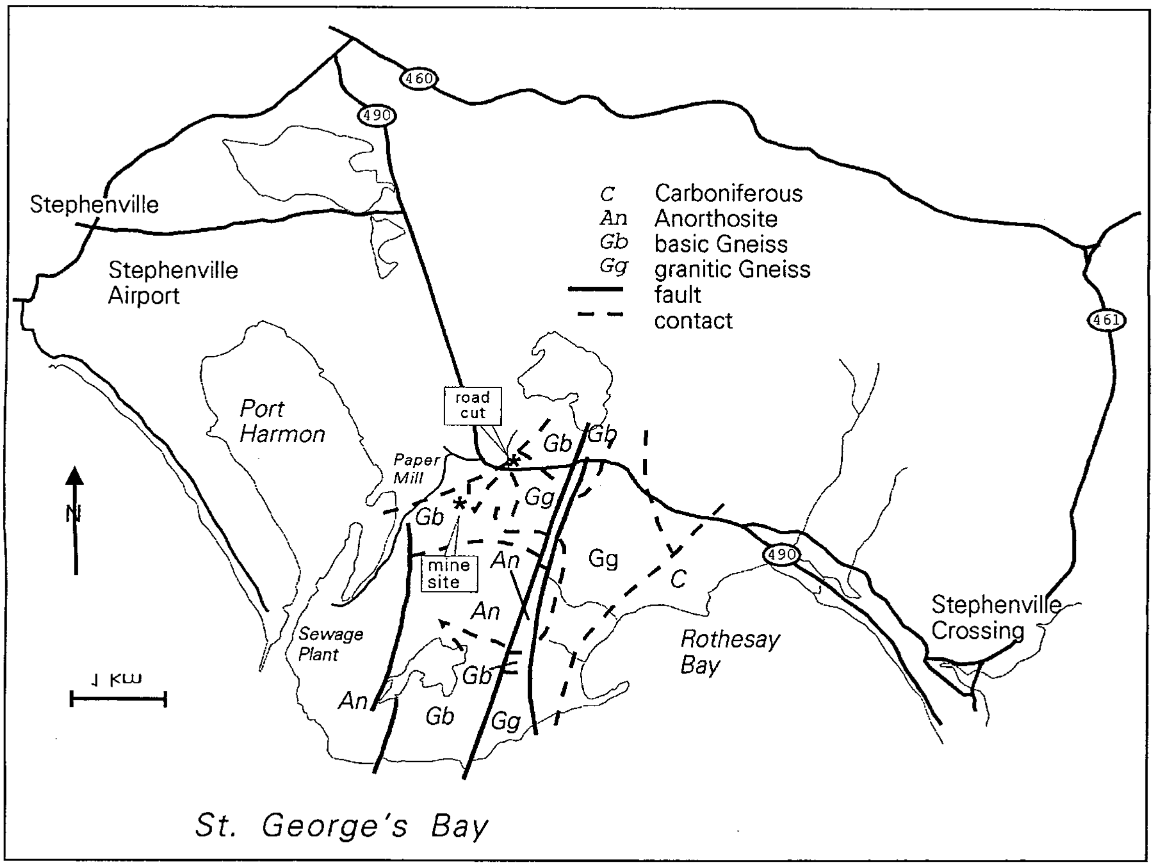

Figure 1: Location map showing regional geology of Indian Head Peninsula. Modified from Murphy and Rao, 1975.

Black Cove and Black Point

Traversing from the Shelf Succession to the Humber Arm Allochthon

Introduction

At this location, on the east coast of Port au Port Bay, west of Stephenville, a complete traverse from the relatively autochthonous shelf succession to the Humber Arm Allochthon is possible. Because the overall strike of the rocks is in places, very nearly parallel to the coast, a complete traverse requires about 5 km of walking along the shore. This account (following James et al 1980) suggests one relatively long traverse, but it is possible to make several shorter visits to different parts of the section. The road from Port au Port East to Point au Mal runs a few hundred metres from the shore, and there are a number of access paths down to the shore. However, many of these provide access to cottages and cabins; *visitors should take care not to cross private property without first seeking permission from landowners*. The traverse is most accessible at low tide. Obviously it is not possible to walk the entire coast exactly at low tide; if you are attempting the complete traverse, you should try to work on a falling tide so as to time your arrival at the northernmost point (Black Point) to coincide with low tide.

CAUTION: some parts of this section are impassable at or near high tide. Although there are numerous paths into the woods from the shoreline, it is possible to get cut off by the tide in some coves. If you have to walk the section while tide is rising, keep an eye on your exit route at all times.

Geological Background

From Middle Cambrian to Early Ordovician time the eastern margin of what is now North America was fringed by a continental shelf on which limestones and dolostones were deposited in a warm, shallow marine environment. The upper part of the shelf succession is represented by the St. George Group. At the end of the Early Ordovician, the shelf underwent a subtle uplift relative to sea level, or sea-level fell, producing a disconformity. Most recent investigators (Jacobi 1980, Knight et al 1989) have inferred that the disconformity resulted from uplift at a peripheral bulge as the continental margin approached a subduction zone which lay to the east (in present-day coordinates). However, the disconformity also approximately coincides with a regional sequence boundary (the Sauk–Tippecanoe boundary) in North American stratigraphy, which may be due to a worldwide fall in sea-level.

Whatever the origin of the disconformity, the shelf rapidly became unstable and subsided rapidly. Initial deposits of the Middle Ordovician Table Head Group, above the unconformity, are of shallow-marine limestones (Table Point Formation), indicating that sedimentation was able to keep pace with subsidence. Eventually, deeper-water limestone of the Table Cove Formation was deposited when carbonate production was unable to keep pace with subsidence. Locally, the Table Cove Formation passes laterally into thick submarine limestone conglomerate (Cape Cormorant Formation), indicating that faults were actively dissecting the former continental shelf.

The Table Head Group is overlain by shales and sandstones and further limestone conglomerates of the Goose Tickle Group, derived from a mass of deformed and thrusted rocks advancing eastward. Eventually, the margin was overthrust by deformed deep-water sedimentary units and ophiolites of the Humber Arm Allochthon. The allochthon includes numerous fault-bounded blocks and slices, many of which were derived from a succession (the Humber Arm Supergroup) covering approximately the same time-interval as the shelf succession, but representing environments found on the continental slope and rise.

Traverse Description

The lower part of the succession may be reached from Highway 460. Park at the east end of The Gravels—the pair of gravel bars joining Port au Port peninsula to the mainland. Proceed eastward along the beach for about 700 m to the first bedrock outcrop (location 1 on Fig. 1).

The first part of the section is in the upper St. George Group (Early Ordovician). The first few outcrops (about 5 m of section) probably represent the uppermost part of the Catoche Formation, which is marked by a massive dolostone overlain by a relatively pure white peloidal limestone (Costa Bay Member of the Catoche Formation). The remainder of the section in the St. George Group (about 60 m) belongs to the overlying Aguathuna Formation. The Aguathuna Formation consists of alternating grey to white limestones and yellowish dolostones. Individual limestone and dolostone units typically range from about half a metre to several metres in thickness. Some of the dolostones are finely laminated algal boundstones (dololaminites) that represent deposition in extreme shallow water conditions—probably intertidal. In places mudcracks indicate subaerial exposure. Exquisitely exposed domed stromatolites, locally with domes up to a metre across, are common in both limestone and dolostone beds. These too are indicative of very shallow water conditions. Other units lack laminations, but instead show a mottled texture, suggesting bioturbation. In some of these units the individual burrows are clearly visible because they have been more completely converted to yellowish dolomite than the surrounding grey matrix.

At location 2, about 200 m north of the first outcrop (immediately below a cliff-top house) the coastal section is cut by northeast-dipping thrust faults. A zone of folding about 5 m thick, in which folded beds dip steeply west, marks the zone of deformation. Similar, but larger, zones of deformation are widespread in the area inland at Table Mountain (Palmer 1995).

About 400 m north of the start of the section, the contact between the St. George and overlying Table Head Group is exposed (location 3). The boundary is at the top of a splintery yellowish-white dolostone. The overlying limestone of the Table Head Group is grey and much more uniform in composition than the underlying Aguathuna Formation. At the contact, two surfaces, separated by about 50 cm, appear to show disconformable relationships; this is consistent with the observations of Knight et al (1989) that the unconformity has a complex structure and is sometimes represented by multiple surfaces.

Continuing north, the Table Head Group is represented by fossiliferous limestones (Table Point Formation) that appear well bedded when viewed from a distance, but which have a rubbly to mottled texture, lacking laminations and other sedimentary structures, when seen close-up. This is almost certainly due to very intense disturbance (bioturbation) by burrowing organisms. Fossils are common in the Table Point Formation. James et al (1980) report *Murchisonia* spp., *Maclurites* spp., *Leperdita.*, trilobites, orthids and strophomenid brachiopods, *Patellispongia*, *Nevadocoelia*, *Stormatocerium*, orthocone cephalopods, *Eridotrypa*, crinoids, and calcareous algae. The Table Point Formation is estimated to be 52 m thick on the Black Cove coast, although the section is covered in places and is locally interrupted by faults. However, the strike is nearly parallel to the shore, so it is necessary to walk nearly a kilometre along the shoreline in order to reach the top of the formation.

The succession at the top of the Table Head Group is exposed on the shore below a garbage dump located about 3 km north of Port au Port East on route 462. At the top of the Table Point Formation (location 4) there is a rapid transition into the thinly bedded 'ribbon' limestones of the Table Cove Formation. Up-section again are black, graptolite bearing shales of the Black Cove Formation (Goose Tickle Group); these are in turn succeeded by interbedded shales and sandstones of the American Tickle Formation, locally intensely deformed by small-scale folds and faults. About 35 m above the base of the American Tickle Formation there is a thick unit of limestone conglomerate, that strikes almost parallel to the coast for over 1 km. This conglomerate unit was formerly correlated with the Cape Cormorant Formation of the Table Head Group (e.g. Klappa et al 1980); however, it is now known to be one of a number of lithologically different conglomerates found within the American Tickle Formation, distinguished as the Daniel's Harbour Member (Stenzel et al 1990).

The upper part of the section may be explored by starting at the north end of the shore-parallel outcrop of the Daniel's Harbour Conglomerate (location 5). The conglomerate is capped by a unit of sand-size transported carbonate particles (a calcarenite) which is in turn overlain by a 80–100 m succession of graded sandstones and shales, representing the bulk of the American Tickle Formation at this locality. At a small cove (location 6), the sandstones and shales are overlain by much more deformed rocks of the Humber Arm Allochthon. While the upper beds of the Goose Tickle group are slightly deformed, the overlying Humber Arm Allochthon at this locality consists of intensely deformed, scaly shale containing a mixture of blocks, both angular and rounded, derived from a variety of sedimentary units. This mélange has a texture and fabric similar to modern trench deposits; it probably formed as units from the continental slope and rise were dragged into a subduction zone that encroached on the continental margin during Middle Ordovician time. Later, during the Acadian Orogeny (Devonian) the allochthon was pushed further over the former continental shelf.

To the north are variably deformed sedimentary rocks and mélanges of the Humber Arm Allochthon. Most of the rocks exposed belong to two formations in the Humber Arm Supergroup. Black shales, and reddish and greenish siliceous shales and cherts are probably correlatives of the Middle Arm Point Formation (named for a location in the Bay of Islands). On the basis of lithological correlation, these are probably of Early Ordovician age, approximately contemporary with the St. George Group of the shelf succession, seen at the start of the traverse. Obviously the Middle Arm Point Formation represents a very different environment from the St. George; it was probably deposited in deep water of the outer continental rise. Interbedded sandstones and shales are also found within the Humber Arm Allochthon. Most of these coarser-grained clastic successions are probably derived from the Eagle Island Sandstone—a unit that records the onset of deformation of the continental margin in the Middle Ordovician. Again, fossil evidence is scarce, but these are probably approximate time equivalents of the Table Point Formation. Thus clastic sediments derived from within the growing orogen had already invaded the deeper-water area (where the Humber Arm Supergroup was deposited), at a time when the continental shelf succession was still receiving carbonate sedimentation.

Structures within the Humber Arm Allochthon are best viewed at Black Point (location 7), where superb cliff-high asymmetric folds are exposed in red and green shales and cherts of the Humber Arm Supergroup. Contrary to expectations, the hinges of these folds trend almost east-west (not northeast-southwest as might be expected from the regional structure of the orogen. Possibly the block that exhibits the folds was more or less "floating" in mélange and were rotated into their present orientation during intense deformation that the Humber Arm Allochthon has undergone. These folds are some of the most photographed in Canada, and have appeared in several textbooks and government publications. For the best photographs, you should walk out on the wave-cut platform at low tide.

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Contributor

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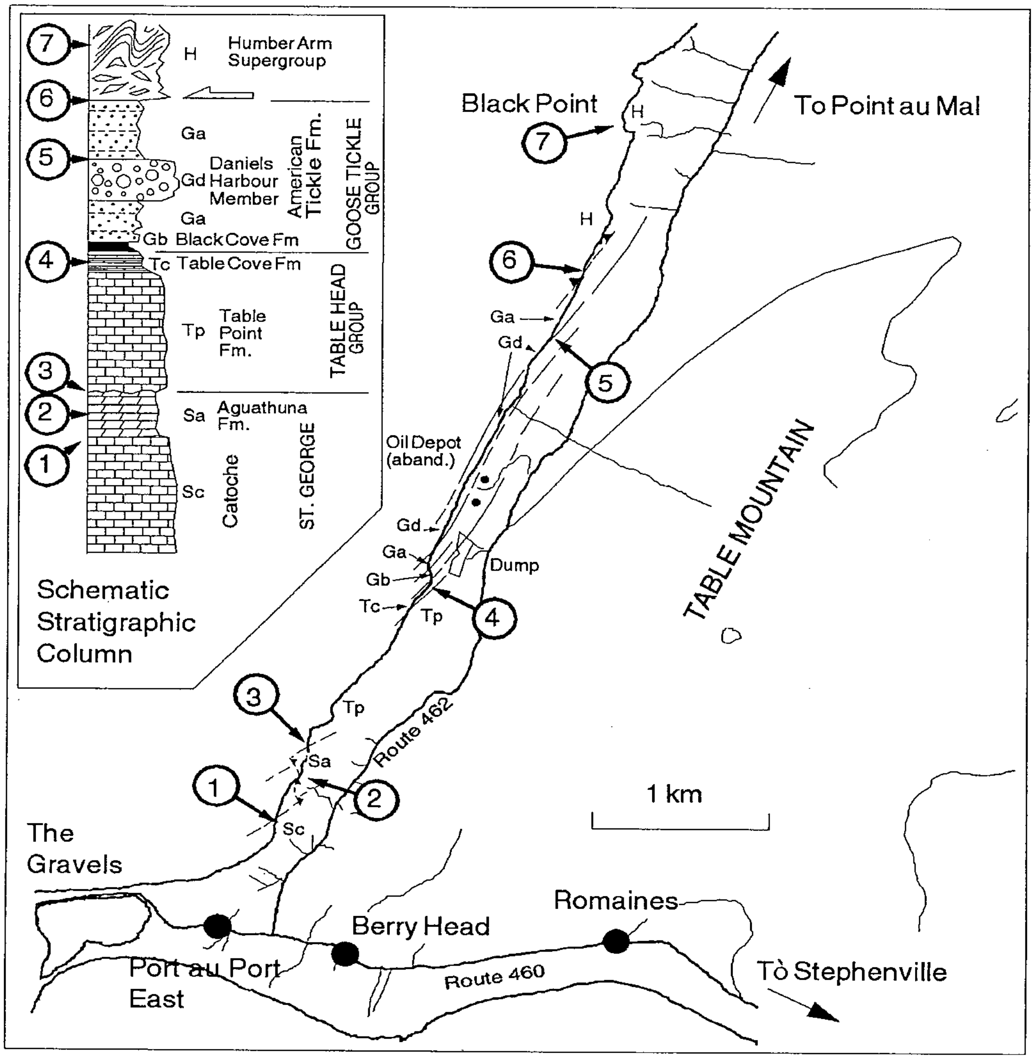


Figure 1: General geology map with schematic stratigraphic section showing locations of stops.

Aguathuna Quarry

Three Ages of Fossiliferous Limestones

Location

Drive west on highway 460 from Stephenville to the Port au Port peninsula. At the Gravels continue directly west onto route 461, the north coast road, and 3 km west from the turn off (and 2 km west of Father Joy's Road) is the quarry. The road goes through the main part of the quarry (Fig. 1). Turn into the quarry and park anywhere convenient.

Introduction

The southern part of the Port au Port Peninsula is underlain by an autochthonous sequence of Cambrian to Middle Ordovician, mainly carbonate, rocks which form prominent bare ridges trending generally east-west. The beds dip gently to moderately north and are offset by a series of northeast-trending high-angle faults.

Northwards, mainly clastic sedimentary rocks of the allochthonous Humber Arm Supergroup are developed around West Bay while an autochthonous sequence of grey micaceous sandstones and shales of Middle Ordovician age occurs in the vicinity of Mainland. The Humber Arm Supergroup is structurally complex with local steeply dipping sequences interrupted by zones of chaotic melange; the Mainland sandstones, on the other hand, dip moderately and uniformly to the northwest, except locally where beds are steeply dipping to northerly overturned.

The long, thin promontory known as Long Point, forming northeast extension of the Port au Port Peninsula, is underlain by carbonates and clastics of the neo-autochthonous Middle Ordovician Long Point Group and the red sandstones, shales and minor limestones of the Late Silurian Clam Bank Formation. The Long Point-Clam Bank sequence dips moderately northwestwards, except to the southwest of Lourdes where the beds are vertical to northerly overturned. The Long Point Group is interpreted to unconformably overlie the allochthonous Humber Arm Supergroup at West Bay.

Carboniferous rocks occur in small, widely separated areas. They consist of flat-lying to gently-dipping coarse conglomerates, plant-bearing sandstones and profusely fossiliferous limestones.

Description

This quarry was operated from 1913 to 1969 with the limestone of the Table Head Formation being removed for metallurgical lime. The rock was shipped to the Sysco Steel Mills in Sydney, Cape Breton Island.

In this one locality the Lower Ordovician, Middle Ordovician, and Mississippian strata, all separated by unconformities, can be examined (Fig. 2). The wall of the quarry displays the upper dolomite beds of the St. George Group (Aguathuna Formation) truncated by an erosional unconformity and overlain by the dark limestones of the Table Head Formation. An 'island' of limestone in the centre of the main quarry (see Fig. 1 for location) is Carboniferous in age (Codroy Group) and was deposited in a post-Ordovician sinkhole. The deposits are all richly fossiliferous.

The Quarry Wall

Here the unconformity separating the Sauk (latest Precambrian to early Ordovician marine incursion and then withdrawal which deposited marine sediments onto the North American continent) and Tippecanoe (Early Ordovician to Early Devonian marine incursion) megasequences is well displayed. Between 15 and 20 m of the upper St. George Group (Aguathuna Formation) is exposed, depending on the degree of downcutting of pre-Table Head erosion. The St. George here consists of cyclic burrow-mottled dolomites, dolomitized algal laminites and stromatolites, tidal flat dolomites, with subtidal limestone beds at the top containing fossils (including *murchisoniid* gastropods). The contact with the overlying Table Head is erosional, with 4 m of relief presently apparent. From east to west along the quarry face, the contact cuts down rapidly, forming a channel-like feature about 50 m across, and then more gradually over 150 m the contact climbs, with two prominent erosional steps. Total relief on the surface in Aguathuna Quarry is 7 m, but relief in the Port au Port area is about 7 m and regionally over 100 m. The possibility of a karst surface at the top of the St. George Group is of some controversy but chert nodules near the top of the St. George appear to be replaced anhydrite nodules as they contain relic sulphate crystallites. In addition to this, two periods of erosion are postulated because local chert and dolostone clast conglomerate has been eroded and illustrates small-scale solution sculpture filled with Table Point limestone. These two surfaces are indicated on the map of the quarry wall (Fig. 2).

During initial transgression over this lithified and eroded surface, fenestral limestones were deposited in the channel. With continued deepening, subtidal mottled and nodular mudstones and wackestones were deposited in the channel, while supratidal mud-cracked conglomeratic algal laminites were being deposited on the elevated surface to the west. During the subtidal conditions in the channel, the deepest part was the site of deposition of a cross-laminated grainstone channel fill. Shallowing once more occurred in the channel and intertidal algal laminites were again deposited, before a second transgressive pulse that gave rise to the more typical Table Head lithologies, exposed in the upper cliffs. The sequence of initial deepening followed by subaerial exposure and karst formation resulting in an unconformity is interpreted to represent uplift related to the peripheral bulge created by loading of the outer part of the shelf with stacked thrust sheets of the Taconic Orogeny. Thus, the return to shallow marine sedimentation for the overlying Table Head Formation probably represents an early phase of foreland basin sedimentation rather than a return to normal continental margin sedimentation.

Initial deposition on the erosion surface is in the form of restricted environment and peritidal limestones, which grade laterally into one another and are eventually buried by subtidal limestones of the Table Head. The mottled limestones of the Table Head illustrate a complex and yet to be deciphered history of diagenesis involving burrowing, partial early lithification, rotation of cemented areas, compaction and dolomitization.

The 'island'

The 'island' (Fig. 2) of brown and buff-coloured Mississippian limestone (of the Codroy Group) in the center of the quarry was not mined because it contains high levels of impurities such as pyrite, galena, barite, and celestite. These rocks are also seen towards the old quarry buildings (see Fig. 1 for other locations of these Carboniferous rocks). This richly fossiliferous limestone is but one of several carbonate buildups that developed in karst valleys, open joints, and other surface irregularities eroded into the underlying Ordovician carbonates during Mississippian time.

The buildups are of variable geometry and composed of the trepostome bryozoan *Stenoporella*? together with algal thrombolites. They partially infill the karst valleys with massive biolithites plastered against valley walls whereas mounds, both isolated and in clusters, occur on valley floors. Together with the well-preserved baffles network of bryozoan/algal structures, there is a prolific but low diversity fauna of brachiopods, molluscs, conularids, and worms.

The following forms can be found: *Dielasma latum*, *Productus (Linoproductus) lyelli*, *Leptodesma acadica*, *L. dawsoni*, *Edmondia rudis*, *Conularia planicostata*, *Serula annulata*, *Martinia galalaea*, *Ambocoelia acadica*, *Spirifer nox*, and *Productus avonensis*. The crustacean *Bellocaris newfoundlandensis* has also been described from this location. Most of these forms are marine in origin. The sediment between skeletons occurs as multigeneration, geopetal infill. Isopachous micritic and fasicular-optic cements and spherulitic calcite cements are common and interpreted also to be marine in origin. The relatively high content of skeletal organisms sets these late Mississippian accumulates apart from contemporaneous structures elsewhere.

The carbonates have also undergone extensive subsequent diagenesis during burial involving fracturing, calcite precipitation, dissolution, sulphate precipitation (celestite and barite), sulphide precipitation (pyrite and galena), and stylolitization. The pockets cannot be collapse features since clasts of the underlying Table Head are not present but must have been original cavities into which the fauna was swept or in which it grew. If the former in true, perhaps they represent solution holes through which nutrient-rich groundwater entered the sea and supported an abundant filter feeding fauna and associated predators. This groundwater flux may also have been the mineralizing agent. This was probably an unsettled, stressed environment as seen in the absence of normal marine biota (e.g., corals, echinoderms, and clacareous algae), high numbers and low species diversity of dominant taxa, and the presence of siliciclastic sequences which indicate influx of freshwater from adjacent uplands (emergent portions of Port au Port Peninsula).

These carbonates are overlain by fluvial sandstones containing plant fossils, and conglomerates.

Source

James, N.P., Klappa, C.F., Skevington, D., and Stevens, R.K., 1980. Cambro-Ordovician of West Newfoundland–Sediments and Fauna; Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Halifax '80; Field Trip 13, 88 p.

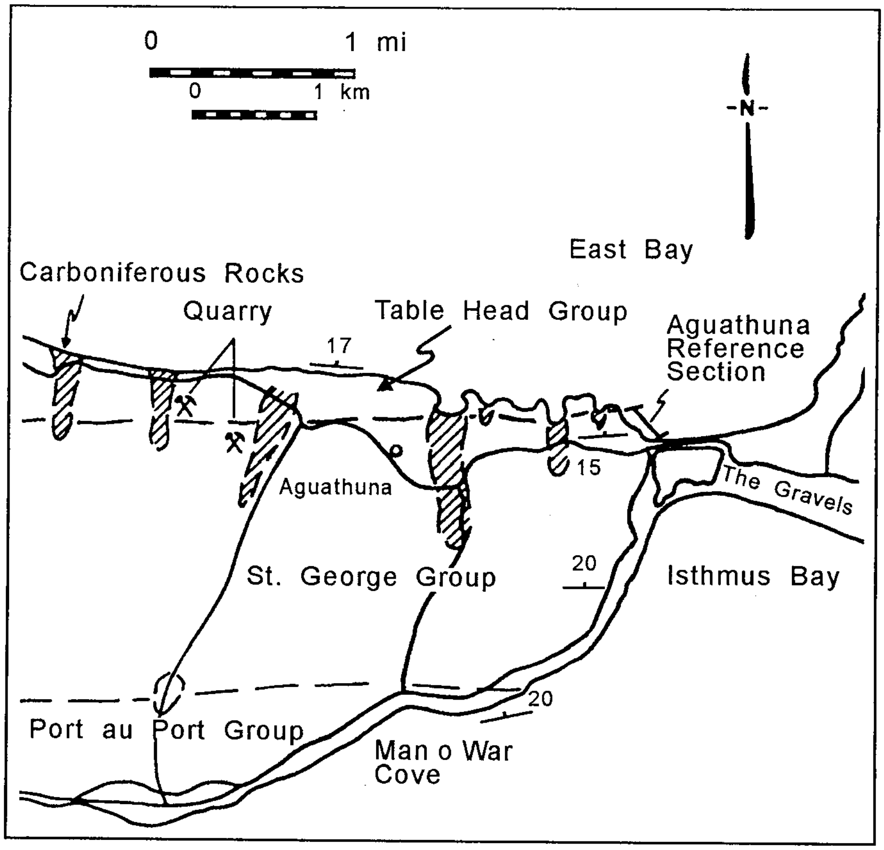


Figure 1: Location map of Aguathuna Quarry showing the regional geology.

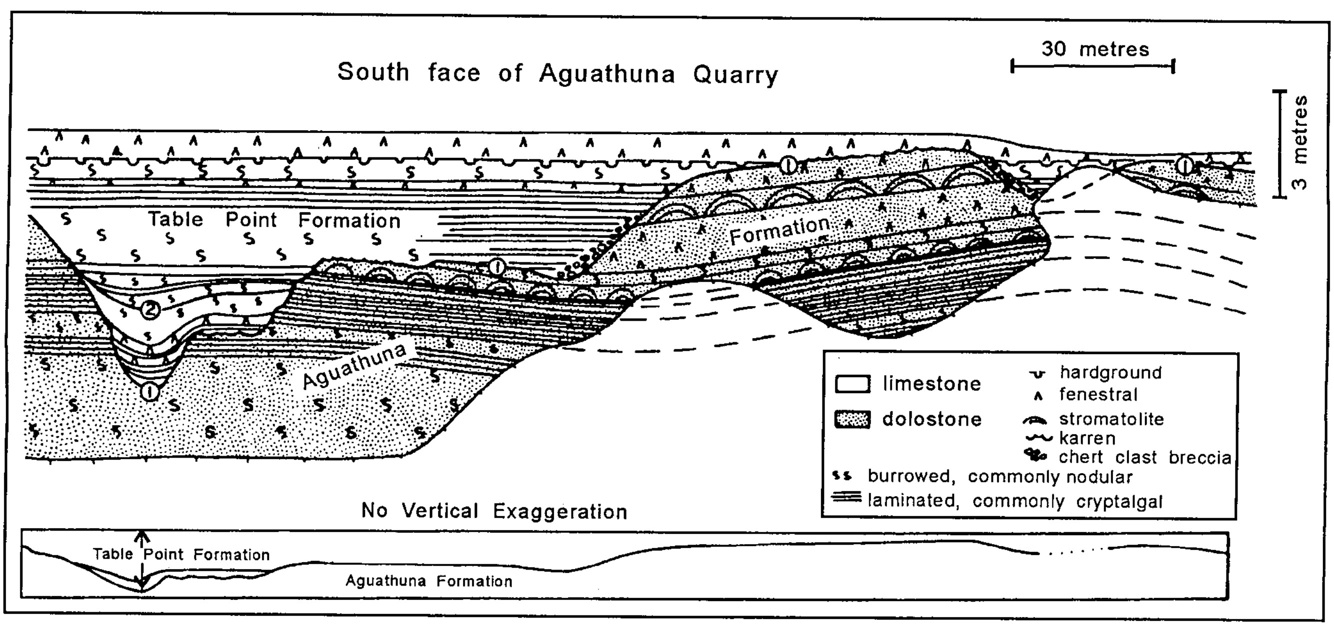
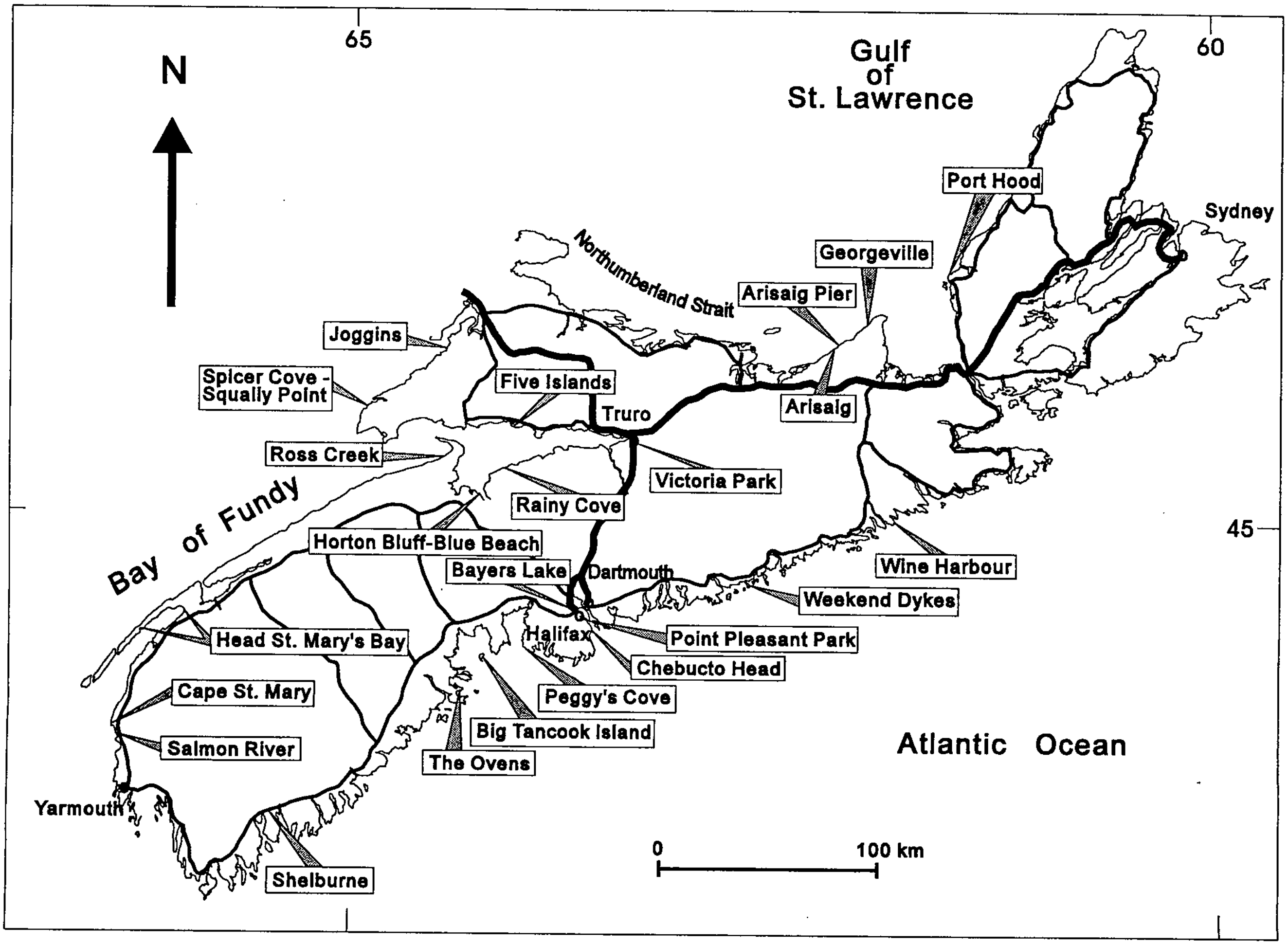


Figure 2: Detailed geology of the south face of the quarry at Aguathuna. From James et al. 1980.

NOVA SCOTIA



Outline map of Nova Scotia showing locations of field trip descriptions.

Port Hood

The Upper Carboniferous Colindale Member Type Section

Highlights

The Colindale Member type section, as defined by Giles et al., (in prep.), is exposed along the western coast of Cape Breton Island approximately 4 km northwest of the community of Port Hood (Fig. 1). Approximately 350 m of continuous Upper Carboniferous strata is well exposed along the coast containing primarily grey sandstone, shale, mudstone, siltstone, and coal, and showing good representation of cyclic sedimentation (Fig. 2).

Location

The section starts at an abandoned sandstone quarry at the end of Route 19 and continues along the coastline southward for approximately 2 km, almost to the Murphy's Pond wharf. To access the base of the section, one must obtain permission from the land owners at the end of the paved portion of Route 19. Beginning the section at the southern portion of the quarry one can walk through the lower part as far as the Cape Linzee sandstone, where except at the lowest tides, it is inaccessible. The remaining part of the section may be accessed from the other end where parking is available just beyond the wharf, on top of the hill.

Stratigraphy

Strata of the Colindale section belong to the Port Hood Formation (late Namurian to Westphalian A) of the Cumberland Group. Norman (1935) defined the Port Hood Formation as consisting of massive sandstone and interbedded red and grey shale. He subdivided the Port Hood Formation into two units, the lower of which consists of massive grey-to-buff sandstone alternating with red shale. The upper unit is predominately grey and consists of an interbedded group of massive and cross-bedded sandstone, arenaceous shale, and carbonaceous shale locally containing coal seams. The Port Hood Formation has recently (Giles et al. 1995) been subdivided into the lower Margaree Member and the upper Colindale Member.

The Margaree Member, exposed at many locations along the west coast of Cape Breton Island, reaches a maximum thickness of 2000 m in its type area near Port Hood, where it underlies the Colindale section. The Margaree Member has been dated as late as late Namurian to Westphalian A. This member contains thick sandstone deposits with intercalated greyish-red and subordinate grey mudstones and siltstones that commonly possess little to no stratification and blocky fracture. Thinly bedded, very fine-grained, greyish-red sandstones generally occur in association with the finer-grained sediments. These sandstones have abundant ripple cross-laminations and are ordinarily less than 2 m thick. Thicker, coarser sandstones, separating the fine-grained intervals, are up to 30 m thick, medium- to coarse-grained with abundant plant debris, and typically multi-storeyed.

The Colindale Member, approximately 350 m thick at Colindale, is conformable with bounding strata. Its contact with the underlying Margaree Member is marked by a transition from red to predominately grey overbank deposits. At the point where Gersib (1979) stopped measuring the section, a major fault exists. This fault cross-cuts the top of the section, removing or displacing an undetermined amount of strata, possibly 200 m thick. Above this fault, exposure is intermittent. The Colindale Member was previously dated as Westphalian A from plant fossils (Bell 1943) and spores (Barss 1967).

The Colindale Member, comprising the coal-bearing strata within the upper third of the Port Hood Formation, consists predominately of fine-grained clastic rocks with intercalated sandstone bodies. The finer-grained rocks include: medium to dark grey and subordinate red mudstones characterized by blocky weathering, slickensides, and root traces; black fossiliferous shales and limestones containing abundant bivalves, ostracods, and fish remains; laminated grey siltstone, and mineable coal seams.

The Port Hood Formation, attaining a thickness of 1400 to 3000 m, unconformably overlies the late Visean to early Namurian Mabou Group of western Cape Breton Island. The Henry Island Formation (formerly included within the Inverness Formation: Norman 1935) conformably overlies the Port Hood Formation. The contact relationship between the Port Hood and Henry Island Formation is visible only in drill core retrieved from Port Hood Island.

Background

The succession of rocks at Colindale is very similar in age and character to the classic Carboniferous section at Joggins, northeastern Nova Scotia. The section differs notably from the Joggins section in that no upright trees are visible in the cliffs at Colindale. The

Colindale section was previously described by Gersib (1979) and Gersib and McCabe (1981), concentrating on the sedimentology of the upper Port Hood Formation. They concluded that the Port Hood Formation was deposited within a large, meandering river cutting across a flood plain. Carroll and others (1972) concluded that the rocks of the upper Port Hood Formation at Colindale suggest a deltaic condition of deposition. The most recent sedimentological study is that of Keighley and Pickerill (1996) who addressed the architecture of major sandstone bodies within the Port Hood Formation, concluding that the sandstones in the upper portion are the product on low-sinuosity fluvial deposition of sandstones in the lower flow regime similar to that shown in Fig 3.

Sedimentary Features

Mine Deposits

Coal seams have provided a long history of coal production in western Cape Breton Island. The main coal seam at Port Hood, the 6-Foot Seam, is approximately 2 m thick and was mined intermittently from 1865 to 1966 (Hacquebard et al. 1989). The 6–Foot Seam, ranked as high volatile B bituminous (Gillis 1975), once cropped out on shore at Port Hood but the coal has since been extracted and is no longer exposed (Hacquebard et al. 1989). Many thin (5–35 cm thick), generally impure, coals occur within the Port Hood region with the thickest seam reaching 70 cm thick. This coal seam lies approximately 20 m above the base of the section and contains old mine workings in the form of timbers.

Carbonaceous shales, common to the Colindale Member, are dark grey to black fissile to brittle, containing thin coal stringers (1–10 mm) and abundant traces of compressed plant debris. There are more than 20 carbonaceous shales in the section ranging in from 9-75 cm thick.

Coals generally originate as peat which forms in swamps on low-lying ground in deltas, alluvial plains, and coastal areas (McCabe 1984). The coals within the Colindale Member are autochthonous, or derived from plant material that accumulated in situ. Evidence for an autochthonous origin for most of these coals includes carbonaceous root traces underlying coals and the presence of seat earths.

Lacustrine—Bay Fill Deposits

Finely laminated siltstones and mudstones, characterised by platy weathering continuous siderite bands, and parallel to ripple cross laminations, represent organic poor lacustrine—bay fill deposits. Fossiliferous shales common to the Colindale Member are greyish-black to black fissile to platy, commonly calcareous, well-stratified, and organic-rich. These shales commonly overlay coals and carbonaceous shales or fossiliferous limestones. The fossiliferous shales range in thickness from 10 to 190 cm, although are generally approximately 30 cm thick. The limestones are greyish black to brownish black, bituminous, and contain an abundance of compressed, flattened bivalves (*Naiadites* and *Curvirimula corvosa* Rogers), ostracods, and annelid worms (*Spirorbis* sp.) giving the rock a wavy-stratified appearance. The limestones occur as resistant ledges within or below fossiliferous shales, ranging from 10 to 100 cm thick. Ostracods and bivalves tend to occupy distinct layers but locally occur together.

The micritic nature of these shales and limestones and the abundance of articulated shells suggests that they were deposited in an environment with little detrital input and lack of strong currents. These shales and limestones likely represent shallow water facies formed by the drowning of swamps and poorly drained flood basins, deep enough to provide an environment amicable to bivalves and ostracods.

Flood Plain Deposits

Crevasse splay and levee deposits form sheets of sandstone and siltstone that occur in association with the fine-grained deposits of the section. They may either fine- or coarsen-upward and possess either an abrupt or a transitional basal contact (lacustrine delta). The siltstones within the Colindale Member are generally medium grey with fine parallel to ripple cross-laminations and may contain sparse plant debris, brown siderite nodules, generally coarsening-upward into sandstone. These sandstones have thin flaggy bedding with shale partings. Other characteristics of these sandstones include mud drapes, desiccation cracks, large advancing cross-sets, ripple cross-stratification (asymmetrical current ripples, climbing ripples), and evidence of vegetation in the form of roots at the top of the unit. The crevasse splays at Colindale include both those built out onto a vegetated flood plain and into a body of water (lake or interdistributary bay) (Gersib and McCabe 1981).

Poorly drained flood plain deposits of the Colindale Member consist of unlaminated mudstones, commonly silty, that are medium to dark grey with common grey mottling, notably where organic material is present. These mudstones are composed of angular to sub-spherical blocky peds with *in situ* carbonaceous root traces and macerated plant debris, and lack any obvious stratification. Characteristically, this lithology possesses concave-up joint sets, desiccation cracks, and small randomly oriented slickensides on the ped surfaces. Siderite nodules (up to 35 cm thick) occur scattered within the mudstones.

Pedoturbation may have removed all original stratification leaving the rock unstratified and basically homogenous. The presence of in situ root traces is one of the most diagnostic features of paleosols. These mudstones, or paleosols, represent poorly drained soils that may have developed in a flood plain environment. The features common to these mudstones suggest that they are, at least in part, vertisols, that formed by the swelling and contracting of expandable clays in soils in response to alternating wet and dry conditions (Mack et al. 1993). Grey mudstones presumably represent drowned ancient soils (paleosols) formed in a flood plain environment or as poorly drained soils.

A small percentage of greyish red mudstones occur within the Colindale Member, attaining thicknesses of approximately 6 m, although generally much less. The red mudstones are much like the grey mudstones in character, breaking into angular blocks (0.5–4 cm diameter), possessing no stratification, and are generally homogenous. These mudstones have concave up and planar joint sets spaced 0.5–10 cm apart that can be traced laterally for many metres. Scattered calcareous nodules (1–10 cm), infilled mudcracks (<50 cm deep), and sparse organic matter are other features noted in the red mudstones. The mudstones may possess drab mottling which suggests a change in pH, perhaps related to seasonal variations. These mudstones likely represent cumulative soils that occur on a well-drained flood plain undergoing seasonal variations.

Heavily rooted claystones that generally underlie coal seams possess no stratification and exhibit crumbly to blocky fracture (seat earths). These clay-coated rocks are light to medium grey, generally mottled, and contain organic matter in the form of plant debris and root traces. The presence of these claystones suggests possible emergence or sub-aerial exposure formed either as exposed soils, later drowned with peat accumulations during transgression, or weathered zones below an accumulating peat, with organic acids from peat contributing to their degradation.

Channels

Channel sandstones within the Colindale Member may be either minor (1–2 m thick) or major (more than 2 m thick). The major channels are generally multi-storied sandstone deposits, 2–20 m thick, generally forming headlands. Erosive surfaces separate storeys and incise underlying flood plain deposits cutting up to 5 m of underlying strata. They are generally massive in appearance, and may be parallel- to cross-bedded. The sandstones are light grey with surfaces weathered light olive grey and are medium- to coarse-grained. Some of these sandstones have local baal lag deposits containing carbonate clasts, macerated plant debris, and logs. Other, less common characteristics of these sandstones include mudchip and siderite intraclasts (up to 4 cm diameter and larger) concentrated in thin bands, plant debris (*Calamites* dominating, also *Lepidodendron* and *Sigillaria*), the presence of pyrite nodules (with reaction halos) and specks of haematite.

The Cape Linzee sandstone is an excellent example of a major channel sandstone deposit. Five metres above the base there is a 1.8 m thick lensoidal lag deposit, composed of carbonate clasts. The storeys within this channel body are variable in thickness separated by erosive surfaces. As the top fines upward, the beds decrease in thickness and contain abundant current ripples.

Sequences

Studies of Colindale Member have revealed repeated coarsening and fining upward sequences on a scale of one to ten metres.

The sequences consist of thin coals and carbonaceous shales that overlay intensely rooted mudstones and claystones. Commonly, the coal seams are overlain by fossiliferous limestones or abrupt-based sandstone bodies and underlaid by highly rooted,dark grey, clay-rich zones suggesting that the coals were derived from *in situ* material. Fossiliferous limestones that blanket the coaly intervals are dark grey to black and are overlain by shales of a similar nature. The shales, well-stratified and calcerous, are overlain by coarser infills of mudstones and siltstones that characteristically contain siderite nodules and bands, and possess plat to conchoidal fracture. These pass up into finely parallel- to cross-laminated siltstones and fine-grained sandstones containing siderite nodules that are often rooted in the upper portion. This coarsening-up succession is reminiscent of a shoaling upward sequence characteristic of a lacustrine bay fill assemblage.

Some sandstone bodies in the section fine upward, passing upward into grey silty mudstones that are highly rooted and possess virtually no stratification (paleosols). These mudstones fine upward, often attaining a claystone consistency as they underlie coal seams.

A number of siltstones and mudstones show evidence of exposure with the presence of *in situ* roots, rhizoconcretions, and occasional red coloring. Submergence is represented with the presence of siderite nodules and bands, ripple marks, trough cross-beds, vivalves, ostracods, and fish scales. The presence of siderite may suggest poorly drained flood plains but not necessarily total submergence. Collectively this evidence suggests that the base (water) level was continuously fluctuating. Sequences within the succession suggest that once vegetation in swamply regions to be preservedas coals and carbonaceous shales. Well-preserved roots underlie the coals. Limestones and shales containing abundant vivalves and ostracodes are further evidence of a base level rise that allowed for coal development. Higher in the sequence it is evident that the base levell gradually fell as indicated by the disappearance of fossils and the presence of siderite nodules withing paleosol-like rocks. There is a general coarsening upward succession at this level where the mudstones grade into siltstones and eventually sandstones. The sandstones, in turn, fine upwards to siltstone and eventually mudstone.

Fossils

The section contains excellent examples of alluvial and flood plain deposits containing coal seams and fossiliferous shales and limestones. Fossils within dark grey to black shales and limestones include well-preserved bivalves (*Naiadites* sp., *Carbonicola*, and *Curvirimula* spp.), ostracods (*Carbonita altilis* and *Helboldtina evelinae*), annelid worms (*Spirorbis carbonarius* Dawson) with sparse plant and fish remains (Belt 1968, Gersib and McCabe 1981).

Track ways of giant arthrodpods (*Arthropleura*), 4.5–5.5 cm across, are present on the bedding plane of a large channel sandstone body approximately 50 m above the Cape Linzee sandstone. One the crevasse splay--levee deposits approximately 100 m below the Cape Linzee sandstone. Keighley and Pickerill (1994) identified *Taenidium baretti* (Bradshaw) which are unwalled or unlined meniscate backfilled burrow.

A fossil skeleton of *Romeriscus periallus*, a limnoscelid reptile, was described from Cape Linzee by Baird and Carroll (1967). This skeleton was retrieved in 1959 from a 1.2 m sandstone body, interpreted by Gerisb (1979) as a crevasse splay, between a 25 cm and a 10 cm coal seam between the base of the Colindale and the Cape Linzee sandstone (Caroll et al., 1972). Other fossils found in sandstone beds occurring within this interval include scales of the crossopterygian fish Rhizodopsis, Braopezia-like footprints of large temnospondylous amphibians, and vertebrate of a large embolomerous apmphibian (Baird and Carroll 1967).

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Contribution

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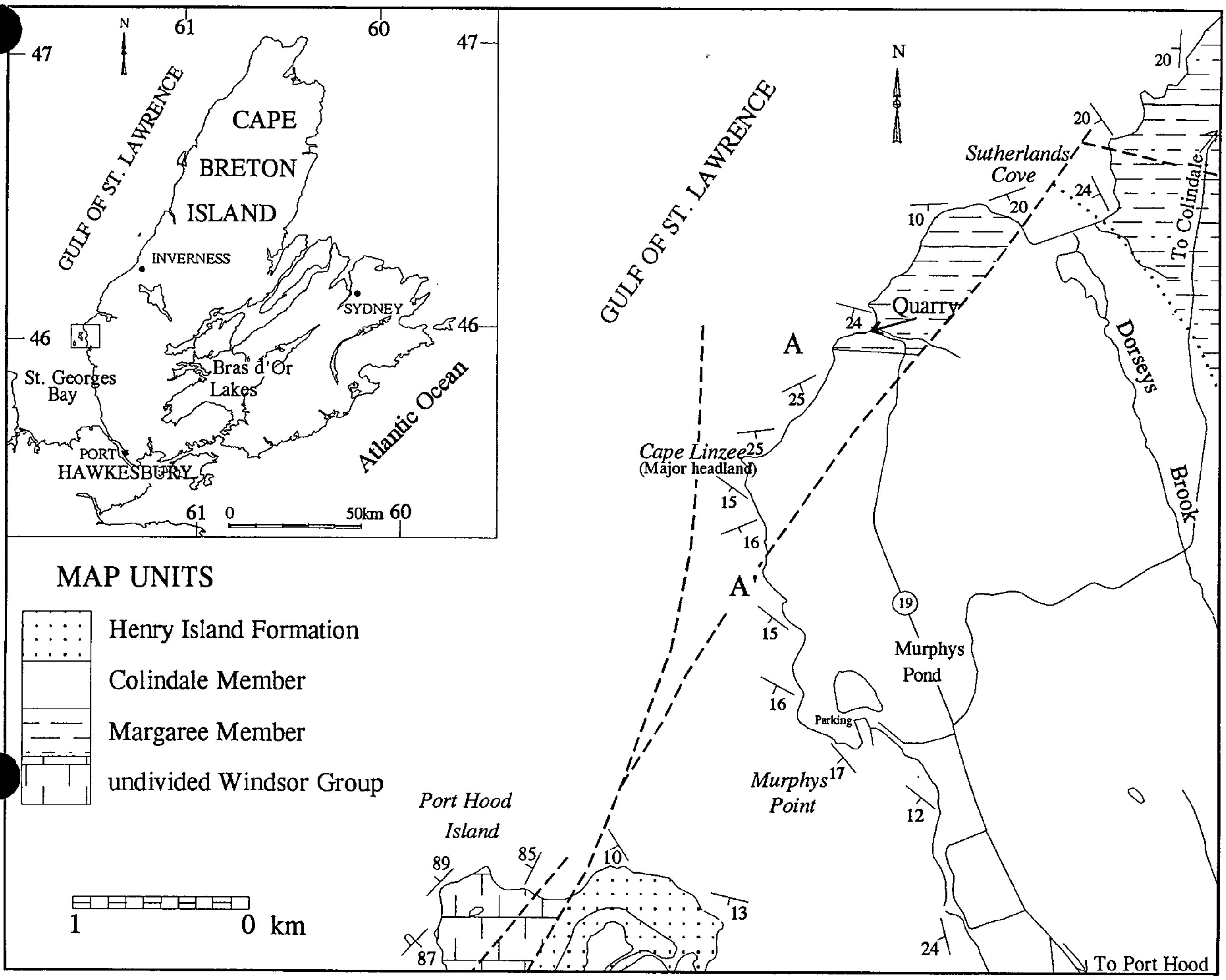

Figure 1: Location map for the Colindale Member type section.

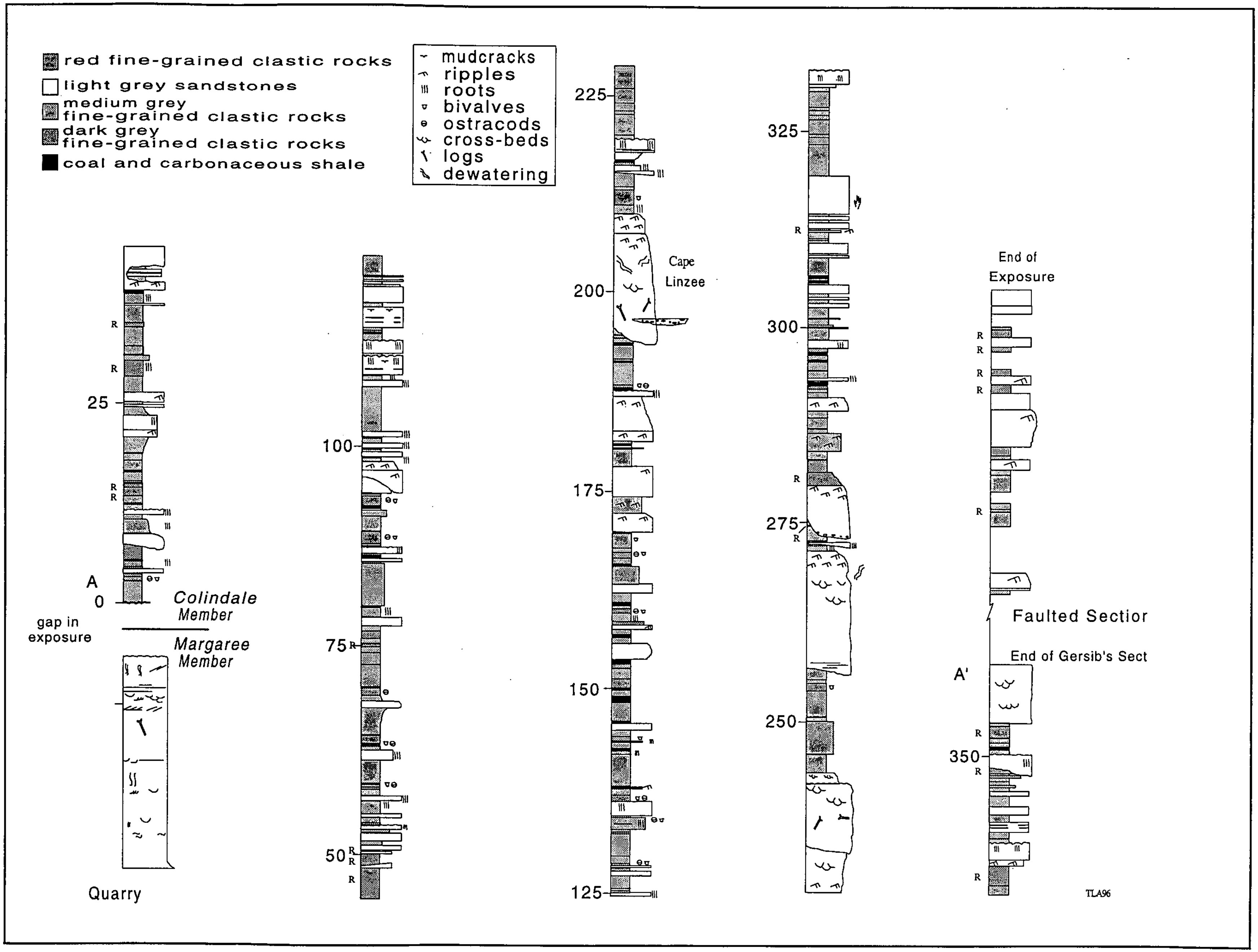

Figure 2: Colindale Type Section

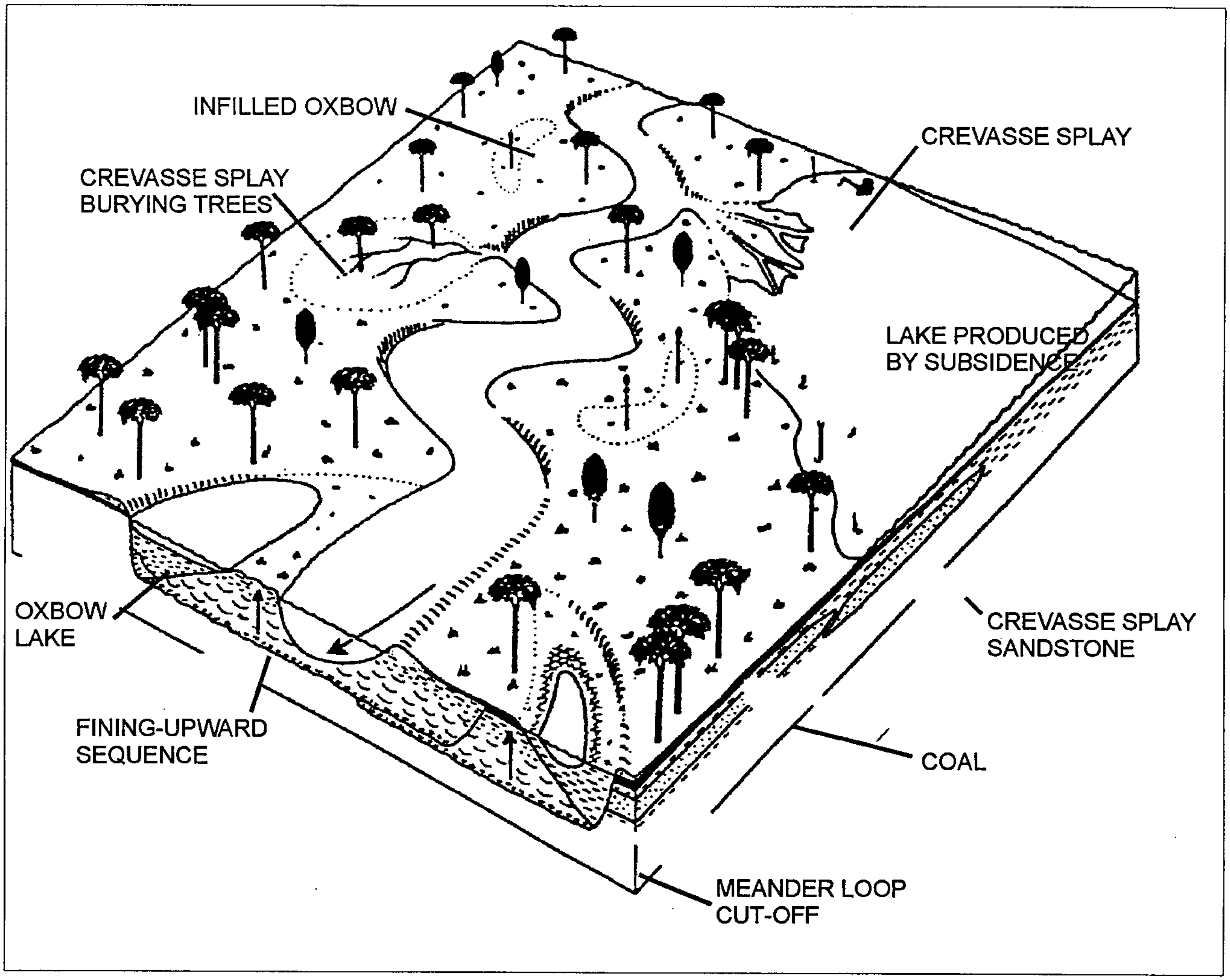

Figure 3: A model for the deposits found in the Port Hood Formation. (Gersib and McCabe, 1981)

Georgeville

Early Cambrian Intrusive Rocks

Highlights

- hornblende gabbro and alaskite granite plutonic rocks
- mafic and felsic pegmatites
- contact metamorphism and skarn deposits
- diabase dykes, quartz and calcite veins

Directions

The trip is along the Northumberland Shore, north of Antigonish on the Cape St. George Peninsula (Fig. 1). It begins at a quarry approximately 5 km northeast of Malignant Cove, off Highway 337, and ends at the Georgeville Pluton, approximately 2 km northeast. Permission and a key for access through the gate and into the quarry may be obtained from C.R. MacDonald Construction Ltd. of Antigonish. Quarry operations were carried out in 1960 by Maritime Rock Products Company. The land along the shore is private.

Special Considerations

This trip is along 2 km of coast containing steep spectacular cliffs, large boulders and deep chasms. Work or hiking boots (not rubber boots) are advisable for hiking the rugged coastline, and a lunch should be taken. Do not jump from boulder to boulder or across chasms and do not attempt to climb steep slopes. Go around. The northern exposure of this coast may require warm clothing, depending on the time of the year the trip is conducted. It would be preferable if the trip to the Georgeville alaskite could be coincident with low tide, otherwise some features may be underwater.

Introduction

The shore of the Northumberland Strait (Fig. 1) provides a unique opportunity to expose students, not only to Geology, but to the related fields of Chemistry and Biology. This guide is suitable for high school Geology or Chemistry, but may be readily adapted to the junior high school program.

Within a very short section of shore, most common geological and geochemical features of igneous rocks are present and represent a spectrum of features that would otherwise require extensive field excursions.

General Geological Setting

The sediments and volcanic flows of the late Pre-Cambrian Georgeville Group (Keppie 1979) form the oldest rocks in the area (Figs. 2 and 3). Benson (1974) interpreted their age as Cambro-Ordovician and the rocks were known as the Brown's Mountain Group. Fletcher (1886) originally considered these rocks as Precambrian in age and subsequent field work by Murphy and Keppie (1987) has shown their structures to be cut by late Precambrian intrusives so they are of definite Precambrian age. They have undergone low grade regional metamorphism (Greenschist Facies) and polyphase deformation.

During the late Precambrian to Cambro-Ordovician time, these basement rocks were intruded by an appinitic (hornblende bearing) gabbroic complex, known as the Greendale Complex, and alaskite (quartz and K-feldspar bearing) granite, known as the Georgeville Pluton (Murphy and Keppie 1987), using the $^{40}Ar/^{39}Ar$ method at between 650-600 Ma whereas the Georgeville alaskite has been dated at 510 $\pm$ 10 Ma from Rb-Sr data (Keppie 1979), 604 $\pm$ 14 Ma from K-Ar on muscovite and 535 $\pm$ 13 Ma from a Rb-Sr whole rock isochron (Murphy and Keppie 1987). Gabbros and diorites are dark coloured due to an abundance of iron and magnesium (ferromagnesium) minerals and are relatively low in silica ("basic" rocks). They crystallize at high temperatures from a magma before the calcium-iron-magnesium poor, light coloured "acid" rock types, such as granite, are formed. As such, the gabbro-diorite-granite complex may be genetically related (Keppie and Murphy 1980).

This means that each igneous intrusive rock type may represent a different phase of crystallization from one parent magma. High temperature minerals such as pyroxene, calcium plagioclase feldspar, and amphibole crystallize first as the magma cools and produce rocks such as gabbro and diorite. The rocks may accumulate at the bottom of the magma chamber as the denser minerals sink in their liquid.

As crystallization continues, the magma becomes depleted in the high temperature precipitate (iron, magnesium, and calcium minerals) and the rocks subsequently formed become progressively more acid in composition. The result of an undisturbed magma undergoing this process, called fractional crystallization, is an intrusion layered from basic rocks on the bottom to more acid rocks on top. MacDonald (1977) proposed this mechanism for the diorite intrusive at Georgeville.

Georgeville Quarry

Outcrop 1

The rock type has been previously described as a diorite (sodium plagioclase feldspar and amphibole) by Benson (1974), Boucot (1974), and MacDonald (1977) but more recently as gabbroic in composition (calcium plagioclase feldspar, pyroxene and amphibole) by Murphy (1979; 1980) and Keppie (1979).

The intrusive stock is chemically very complex and many basic and ultrabasic igneous rock types have been identified (MacDonald 1977; Murphy 1979, 1980; Keppie and Murphy 1980). This gradational variation in composition has led to the speculation that fractional crystallization of one parent magma may have been the mechanism involved in the evolution of the rock complex.

Marble, in some cases with pyrite, is found in the quarry along with the gabbro/diorite. The hot magma most likely metamorphosed xenoliths of limestone that were incorporated into it during emplacement into the Georgeville Group rocks.

Complex chemical reactions between the magma and associated fluid, and the host rocks, resulted in many hydrothermal alteration products (metasomatism). On the gabbro/diorite-host interface, serpentine and pyrite are commonly found. The serpentine varies from a soft, dark green character to a more compact, greenish-yellow product found in lenses in the quarry and northeast along the beach. Slickenside surfaces are commonly found in the serpentine. Both the marble and the soapstone-like serpentine may be carved and polished into very attractive products. Accessory minerals found in minor and/or trace amounts in the diorite complex, include galena, sphalerite, pyrrhotite, chalcopyrite, azurite, ilmenite, and rutile (Bourque 1980).

Outcrop 2

About 100 m along the shore, southwest of the quarry (Fig. 2), a pegmatite phase of the gabbroic intrusion may be observed. Most gabbroic intrusions, as is the bulk of this one, are medium grained. However, pegmatitic phases are sometimes formed although much more infrequently than the acid pegmatites (as is the case of the alaskite). Hornblende crystals 15-20 cm in length are abundant at this location and many are hollow, reflecting a rapid growth rate in a highly volatile magma.

The alignment of the hornblende crystals in this rock has been interpreted as a preferred growth direction in response to regional tectonic stresses. In this case, emplacement of the pluton would take place during an orogenic event, a period of regional deformation and mountain building, such as with a subduction zone (Boucot 1974). However, more recently, Keppie and Murphy (1980) placed the evolution of the intrusive complex between major orogenic events. Preferred orientation of the crystals may then be due to gravitational settling in a magma chamber.

Northeast along the beach, highly brecciated areas with variable rock types provide further evidence of forceful emplacement and complex chemical interactions.

Since the minerals of the more extensive gabbroic intrusion crystallize at a much higher temperature than those of the late phase granitic minerals, the gabbro/diorite probably provided the bulk of the heat for the transformation of the Georgeville Group mudstones and shales into hornfels by contact metamorphism.

Georgeville alaskite

General Features

At this location (Fig. 3), a white alaskite granite composed predominantly of quartz and potassium feldspar is observed in contact with the very dark host rock. The intrusive igneous activity in the area has baked the pre-existing mudstones and shales into hornfels through contact metamorphism. A black diabase dike cuts through the alaskite perpendicular to the shoreline. Diabase is a fine-grained equivalent of gabbro and as such is composed predominantly of calcium rich plagioclase feldspar and pyroxene.

To the extreme northeast of the pluton, along the shore, is found the pegmatite phase of the alaskite. Faults may be traced by offsets in the dyke.

Outcrop 1

This outcrop is the southern area on Fig. 3, about halfway through the alaskite. Look over the whole outcrop before making detailed observations.

The red boundary between the diabase dike and the alaskite prominently shows the effects of contact metamorphism. The intensity of the red alaskite colour decreases with distance from the dike. This "bake zone" is the result of a strong chemical gradient seeking equilibrium. Iron from the hot, basic magma was transferred into the nearly iron-free alaskite by interstitial ion transfer or by infusion of iron rich magmatic fluids (metasomatism).

Other evidence of post-alaskite dike intrusion is found in the grain size of the diabase crystals, which are larger toward the centre of the dike where cooling was more gradual. At the cold alaskite contact the diabase magma was chilled very quickly and resulted in rapid crystal nucleation that was too fast for large grain development.

Xenoliths provide evidence of forceful injection of the dyke material into the alaskite. As the magma pushed its way up through the host rock, blocks of alaskite were broken off and incorporated into the body of the dike (magmatic stoping). The blocks show the same metasomatic effects as the main contact zone.

Right lateral faults nearly parallel to the shoreline are easily discerned by the displacement of the diabase dike up to two metres (Fig. 3). Left lateral faults nearly perpendicular to the shoreline offset up to two metres (Fig. 3) both the previous faults as well as the dike rocks. Some of these faults have been subsequently infilled by calcite.

Outcrop 2

Approximately 150 m northeast along the shore from outcrop 1 (Fig. 3), the pegmatite phase of the granite intrusion is exposed. This very late stage of magmatic crystallization was extremely rich in volatile fluids and elements incompatible with earlier phases of crystallization. As a result, crystallization took place in a relatively spacious environment that allowed crystal growth to enormous proportions. Quartz crystals with diameters of 50 cm or more are common but very well formed hexagonal crystals are found only up to about 10 cm in diameter. Potassium feldspar constitutes the bulk of the remainder of the pegmatite but minor amounts of chlorite are present.

On the west side of the large pegmatitic outcrop as well as near to the water line at outcrop 2 (see Fig. 3) is a light green coloured dyke. The colour is attributable to an accessory mineral called amazonite, a green microcline feldspar. This exotic mineral, prized as a semi-precious gemstone, may be the result of "incompatible" elements concentrated in the late stages of the melt.

On the southwest side of this outcrop, adjacent to the cliff, a very sharp faulted contact between the pegmatite and the hornfels may be seen (Fig. 3). The brecciated nature of the rocks in this zone provide evidence of the shearing forces involved in these minor faults.

Extending southwest into the hornfels, alaskitic dikes may be followed. These dikes have the same composition as the main pluton but are fine to medium grained in texture due to a much more restricted crystallization space.

Outcrop 3

A few metres southwest of the pegmatite, the hornfels forms a prominent structure (Fig. 3). The original shales or mudstones were baked into a very hard, compact, siliceous rock by the intense heat associated with the intrusive igneous activity in the area. The gabbro/diorite (Georegeville quarry) probably accounts for much of this contact metamorphism as it is much more extensive than the alaskite, and forms at a much higher temperature than the late stage alaskite. (see General Geological Setting.)

As contact metamorphism does not involve regional compressional forces, deformation of the rocks did not occur and original bedding in the shale is preserved. This feature may be seen as fine, alternating, light and dark bands nearly perpendicular to the shoreline. Sheared rocks due to faulting are also present at this outcrop.

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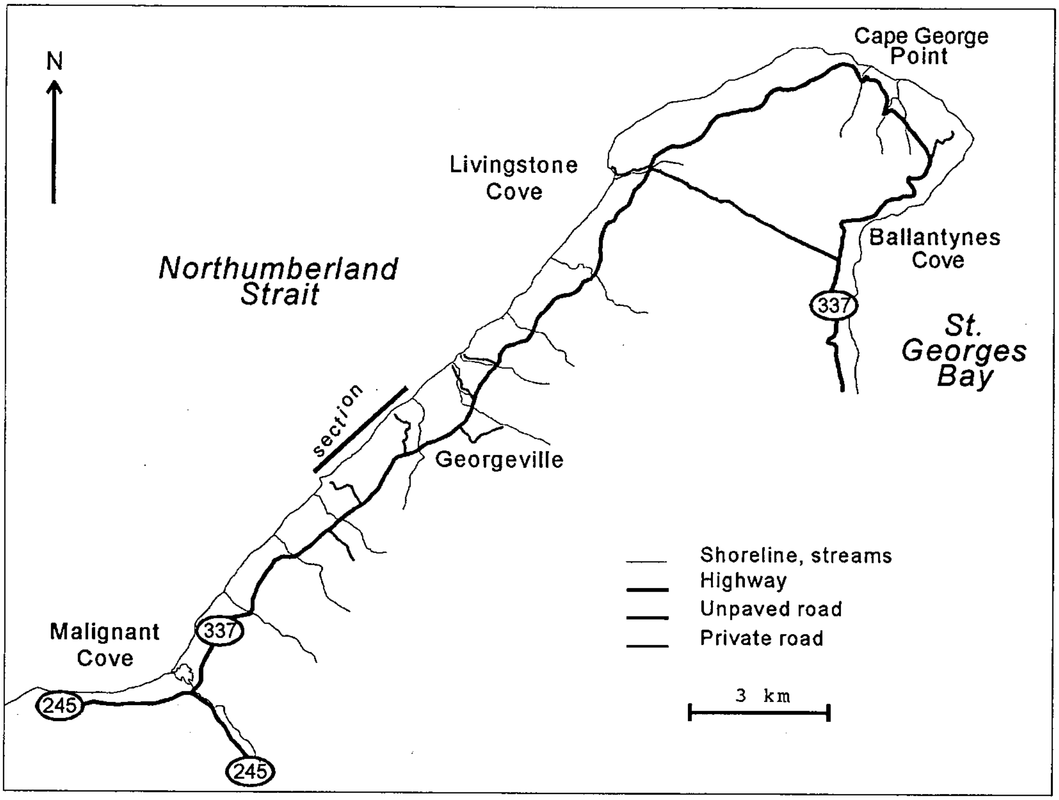

Figure 1: Location map of the Georgeville section

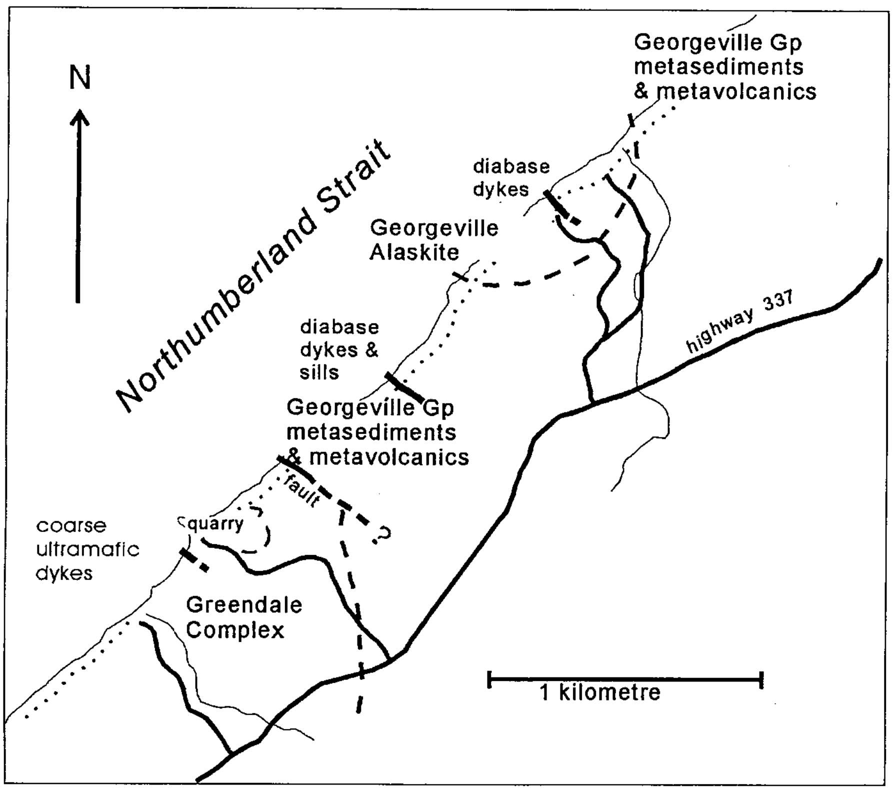

Figure 2: Geology map of the area from the Georgeville Pluton to the quarry, Georgeville, Nova Scotia

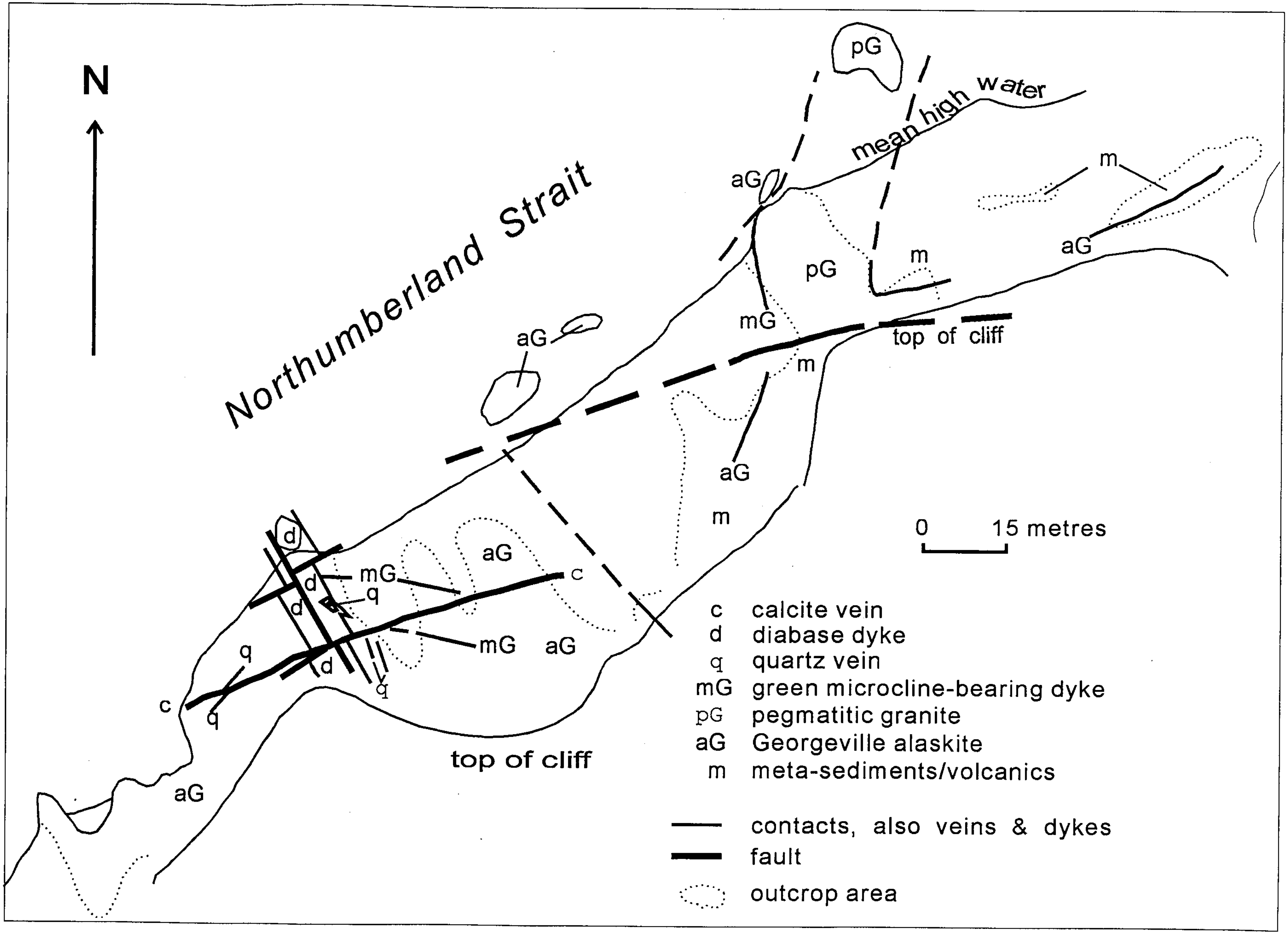

Figure 3: Detailed geology map of the Georgeville Pluton

Arisaig Pier

Rocks of the Silurian Aged Bear Brook Volcanic Group

Highlights

- rhyolite flows, ash fall tuffs and welded tuffs
- basalt-andesite lava flows
- weathering and laterite horizons in mafic rocks
- angular unconformity

Precautions

This is a northern coastal exposure and you may require warm clothing. If the trip coincides with low tide, more features are visible but it is not a prerequisite.

Directions

Take highway 245 east from Sutherlands River (east of New Glasgow; exit 27 off Highway 104) or west from Antigonish to the village of Arisaig on the Northumberland Shore. At Arisaig follow the wharf road to the rock outcrop beyond the fish plant (Fig. 1).

Introduction

The Bear Brook Volcanic Group at Arisaig Pier was extruded onto the eroded surface of the older Hadrynian-Cambrian rocks of the Antigonish Highlands but this unconformity is not exposed at this location. The volcanics, in turn, are unconformably overlain by the Beechhill Cove Formation, Arisaig Group, determined by fossil evidence to be Silurian aged. The radiometric age of the volcanics is 430 ± 15 Ma which places it solidly in the Ordovician. The volcanic rocks here consist of two main compositional types, basalt-andesite and rhyolite. The mafic component forms a pile of discrete flows best seen near Arisaig Pier but occur along the whole shore. The felsic component forms a number of different extrusive types, including flows and domes, ash fall tuffs and welded ash fall tuffs. In addition to these, there are exotic forms exposed further along the shore (e.g., laterite slump sheet).

Basic magma is low in SiO_2 and high in MgO and FeO and, as a result, it is very fluid and can flow for long distances from a vent or fissure after eruption. There are two types of flows, the pahoehoe fluid type of flow and the more viscous aa type of flow. A typical basaltic lava regardless of type has the following characteristics: at the base is a thin layer of clinker balls, spherical to pipe shaped vesicles (some which may curve and hence show flow direction), minor material from the layers below (seen as mud sphericles, etc.), and incorporated lava toes; solid lava with columnar jointing occupying the middle two thirds of the flow; and followed by the top third which is vesiculated lava with clinker, etc. Depending of the time interval between flows and the climate there might be an extensive weathered horizon known as laterite between individual or sets of flows.

By contrast, acidic magmas are high in SiO_2 and low in MgO and FeO and, as a result, are very viscous and have many different eruptive styles. Viscous felsic magmas may be extruded as (1) sticky, toothpasty, degassed *rhyolite* flows or domes; (2) solid fragments and liquid droplets exploded into the atmosphere, which then accumulate near the vent as air-fall tuffs; or (3) dense incandescent clouds of solid and liquid particles supported by expanding gasses (called a *nuee ardente*) which rushed downslope to accumulate as welded tuffs or ignimbrites. Each of these eruptive styles develops an assemblage of textural characteristics, not all of which are unique which makes it difficult to distinguish between the various styles.

Description

Basalt-Andesite: To the northeast of the pier at Arisaig Point and southeast of the fish plant along the shore a short ways (Fig. 2) are exposed several lava flows. They are fine to very fined grained, black coloured, and some contain millimetre-scale phenocrysts of plagioclase and a black mineral now chlorite but probably originally olivine or pyroxene. Most flows are no more than 2–3 m thick, continuous across the exposure area for several tens of metres with the exception of one flow on the beach which terminates within the pile, variably vesiculated and/or amygdaloidal at their tops, and several of the thicker ones show incipient columnar joints in the massive middle portion. Most are separated from each other by variably thick and sinuous red lateritic horizons weathered and eroded more deeply than the flows themselves. Some of the exposures exhibit hydrothermal alteration halos symmetrically disposed along cracks and joints. Much of this is seen as symmetrical bands of discolouration, mineral alteration (sericitization), and oxide mineral (haematite and/or pyrolusite) deposition.

The amygdules contain mainly quartz, chlorite, and prehnite mineralogy with some of the quartz minerals pseudomorphs after zeolites. The flows dip moderately to the north assuming the laterite horizons were original nearly horizontal planes and the concentration of vesicles and amygdules was near the tops of the flows. The contact between the basalt-andesite and the rhyolite is not exposed except for a cryptic zone on the cliff southeast of the fish plant where it could be either interpreted as faulted or conformable.

Rhyolite: At Arisaig Point, for a short distance along the beach northeast of the fish plant, at the knob just east of the wharf road and east beyond the beach are exposures of the rhyolite (Fig. 2). In the wharf area the rhyolite is typically flow banded and porphyritic with fine grained (< 1 mm) phenocrysts of quartz and feldspar crystals. The banding is a series of planar to folded alternating layers of light and dark material which results from layers of minor elemental differences separating during flow due to high viscosity. There is no consistent orientation to the banding or the folds. Within the flows are zones of rhyolite breccia in areas of several metres to tens of metres square, the best exposed at Arisaig Point near to the bottom of the flow (see Fig. 2). East beyond the beach are two rhyolite units, the older one most likely a flow and the younger one pyroclastic. The flow is pink to orange coloured, variably flow banded to massive vitreous up to 30 m thick and very similar to that seen at Arisaig Point. The younger rhyolite is a mottled brown to yellowish to green air-fall tuff grading into welded tuff up to 40 m thick. The air-fall tuff portion has many angular volcanic (up to 3 cm diameter) and mineral fragments, grading quickly from exotic pieces near the bottom to monolithic fragments within the main body. This part of the tuff is generally very heterogeneous in both texture and weathering characteristics. The upper portion of the tuff is composed of volcanic material (*lapilli*) flattened and aligned by compression (called *fiamme*) in an extremely fine grained matrix (originally glass?).

The two rhyolite units are separated by a 3 m thick, red diamictite interpreted to be a laterite slump sheet (Keppie et al. 1978). The diamictite is a pebble to cobble mudstone with rounded clasts, some showing alteration rims and internal flow banding. The matrix is mud-sized. It weathers and crumbles easily but fresh samples show it to have subconcoidal fracture. This unit is the most consistent and laterally continuous rock type along the shore and was used by Zeigler (1959) and Boucot et al. (1974) in correlations and delineating structures.

The Arisaig Group sediments: Unconformably overlying the rhyolites on the south side of the hill east of the wharf road and east of the beach are the basal units of the Beechhill Formation, Arisaig Group. At the wharf road the basal sediments are mainly a yellow, poorly bedded, weakly indurated rhyolite cobble to pebble conglomerate up to 4 metres thick. Clasts are well rounded, in places the larger clasts touch each other, and are composed of flow banded and vitreous rhyolite derived from the rock immediately underneath. Bedding thickness ranges up to 30 cm and becomes more prominent up section along with a grain size decrease. The contact of this unit with the rhyolite is rather uneven, next to the road it is gradational whereas further away into the bush are reddish argillaceous sandy beds and conglomerate portions occupying an erosional scour or depression. Above the conglomerate are a series of interbedded argillaceous sandstone and siltstones and shale.

In contrast to this, on the beach the basal sediments are reddish to pink argillaceous sandstone, siltstone and shale with minor pebble conglomeratic sandstone. The contact is clearly exposed with the sediment resting on the erosional surface of the welded tuff, in some areas the tuff appears to exhibit exfoliation. Clasts are rounded to well-rounded as at the hill and are of the same colour and composition as the tuff.

These sediments were probably deposited in a fluvial environment on the sides or slopes of the rhyolitic deposits and then the advancing Silurian seas modified them. It is very likely that waves produced the well-rounded clasts and weathered some of them from red to yellow colour.

Veins and Faults: Quartz and quartz-epidote veins occur throughout the area and range in thickness from 0.5 mm to 5 cm and in zones up to 1 m wide. Many of the veins show fibrous growth and/or are on minor fault surfaces with slickensided contacts. In the area of the unconformity at the hill are discontinuous veins of barite and agalmatolite (a soft, green mineral used in carving).

Faulting is very prominent throughout the area with horizontal displacement normal to the shoreline in the order of 50 m or less with the notable exception of a larger displacement east of Arisaig Point (see Fig. 2). Many faults can be seen in outcrop as offsets of structures along zones of brecciation and crushing several cm wide and inferred by offsets of major lithologies. Most are the youngest event in the area and are probably related to the Acadian Orogeny or later events.

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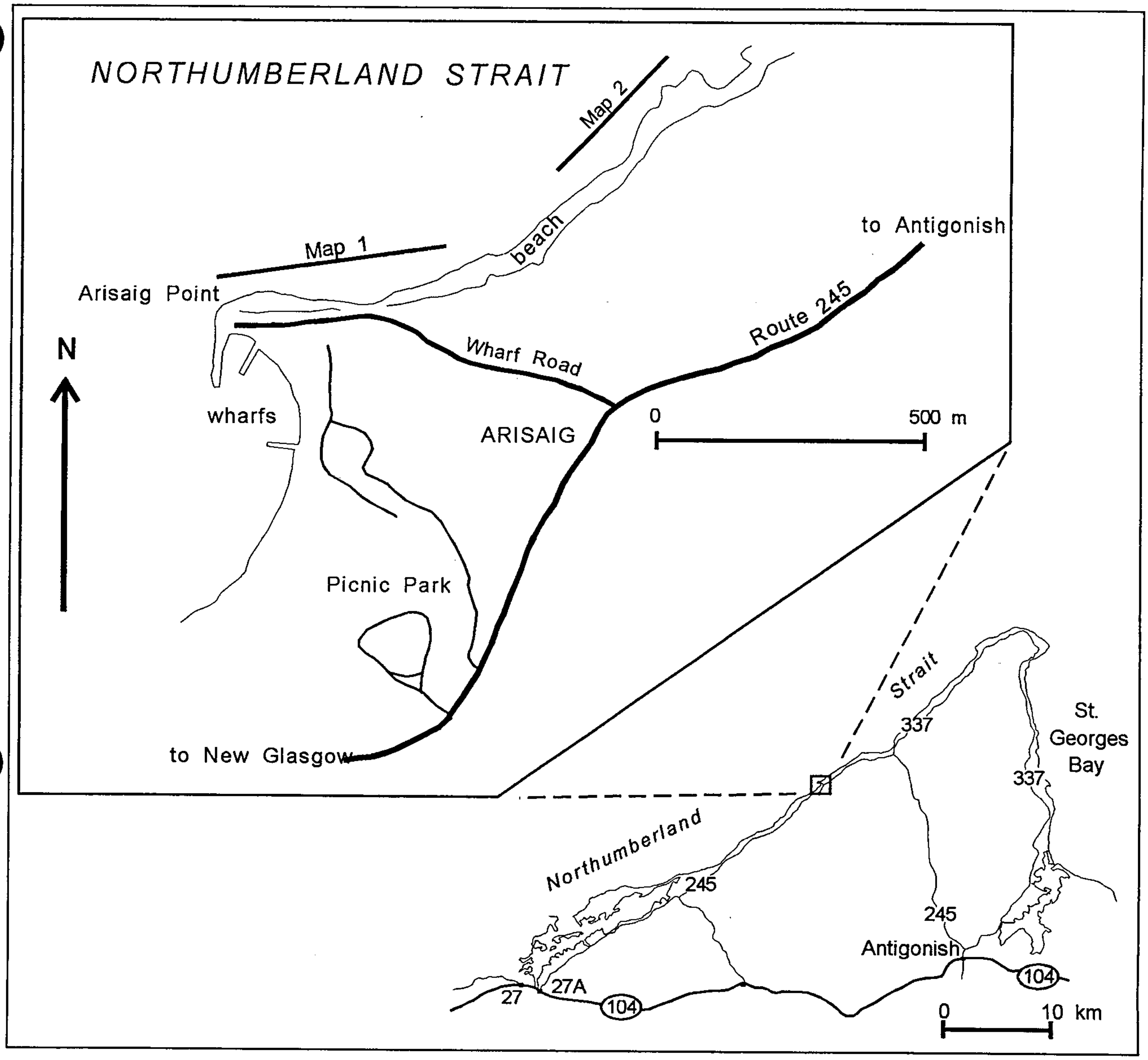

Figure 1: Location map showing locations of maps in Figure 2

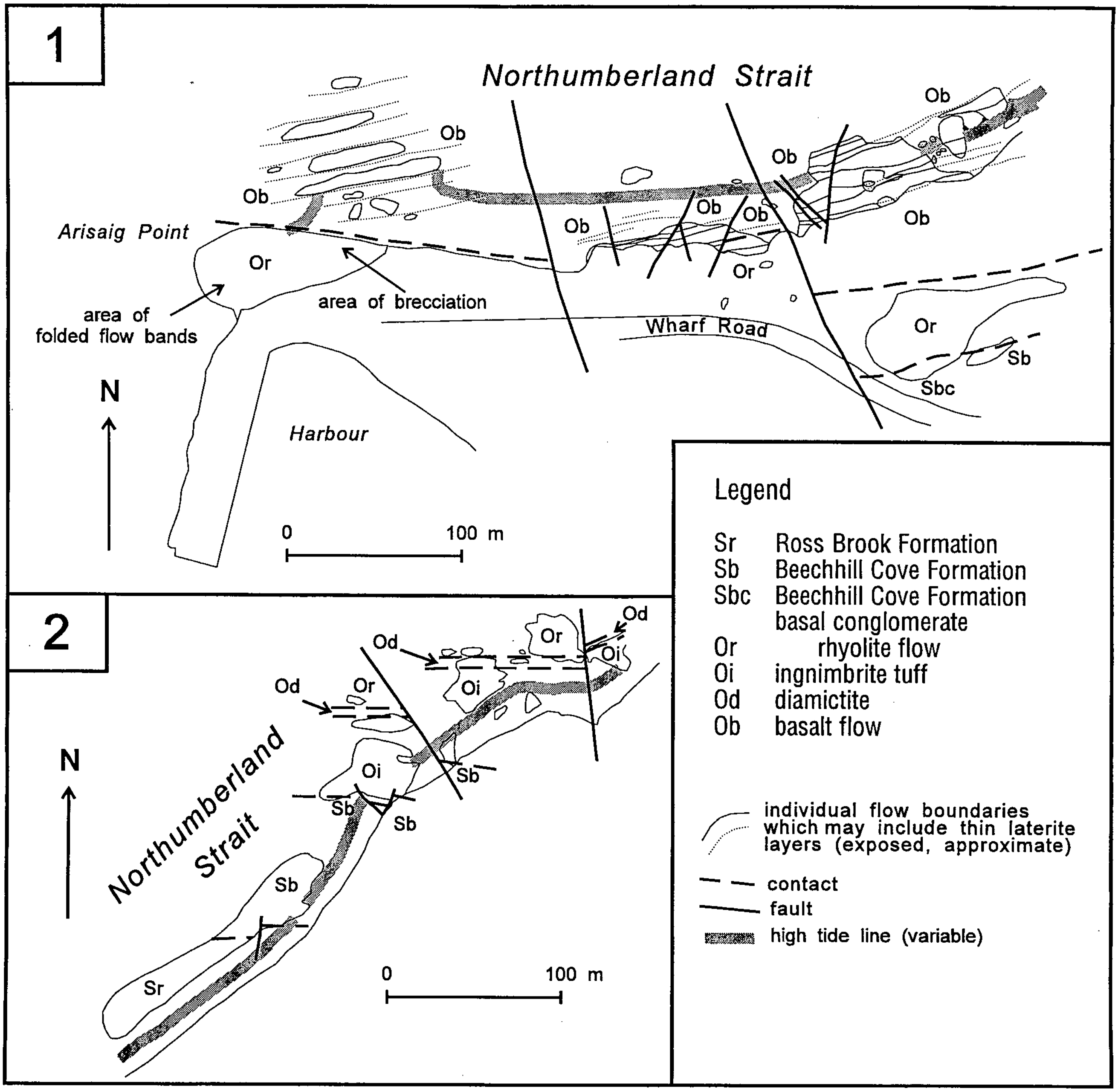

Figure 2: Detailed geology map of Arisaig Pier. 1. - Rocks at the pier and the immediate vicinity to the east. 2 - Shore to the northeast of the pier. Modified from Boucot et al. 1974.

Arisaig

Sediments and Fossils of the Ross Brook Formation, Arisaig Group

Highlights

Silurian-aged marine fauna fossils

Directions

Take Highway 245 east from Sutherlands River (east of New Glasgow; Exit 27 off Highway 104) or west from Antigonish (Fig. 1). At the village of Arisaig, park in the provincial picnic park 500 m west of the wharf road. Walk down to the shore.

Precautions

This is a north facing shore section and as such can get quite cold, depending on the time of year, so dress appropriately. The whole length of beach is faced by steep, unstable cliffs. Do not attempt to climb the cliffs and do not walk along the top; do not stand underneath the cliffs, and good rule of thumb is to stay back as far from the cliff as it is high. This is a protected site so no samples can be collected from the cliffs. Look at the cliffs from afar and only collect beach samples. The tide does not pose a problem here but at high tide a majority of the beach disappears.

Introduction

The Arisaig Group is a continuous sedimentary succession, 1400–1500 m thick, and ranging in age from earliest Silurian (early Llandovery) to early Devonian (Pridoli-Gedinnian). It nonconformably overlies the Ordovician aged Bears Brook volcanics and is in turn unconformably overlain by the red beds of the earliest Carboniferous aged McAras Brook Formation. The Arisaig Group has been divided into eight formations by Boucot et al. (1974) and this trip will be in the Ross Brook Formation. The Ross Brook Formation conformably overlies the Beechhill Cove Formation and is in turn conformably overlain by the French River Formation.

The Arisaig Group has been studied systematically starting with McLearn (1924) and more recently by Hurst and Pickerill (1986). In between there were many workers concentrating on systematic palaeontology (McLearn 1924; Bambach 1969; etc.), selective palaeontology (Peel 1978; Pickerill and Hurst 1986; etc.), and sedimentology (Lane and Jensen 1975; Cant 1980). The reader is referred to these papers for detailed descriptions and locations of particular fauna and rock types.

Description

The Ross Brook Formation is a 280 m thick coarsening-upward succession of interbedded thin fine- to coarse-grained siltstone, dark shale, and bioturbated shale that contain a definite marine fauna. Boucot et al. (1974) divided the sequence into three members based on the siltstone distribution but Hurst and Pickerill (1986) state that the boundaries are essentially arbitrary and cannot be recognized outside the coastal area. In the terminology of Boucot et al. (1974) only the middle and upper members are exposed in the Arisaig Brook area.

Lane and Jensen (1975) provide the best description of this rock, as follows:

> The lower member is 25 m thick and consists of fissile, micro-laminated black shale with graptolites and rare brachiopods. The middle member is a 180 m thick transition zone between the lower and upper members. The upper 90 m of the middle member consists of greenish grey mudstones with thin intercalations (0.5 to 1 cm thick) of siltstone (siltstone/mudstone ratio approximately 1/10, Fig. 12 [sic]). The mudstones commonly have a massive appearance and contain many burrowing structures. These rocks contain a varied fauna, the most diagnostic of which is the brachiopod Eocoelia (Fig. 10 [sic]). This fauna has been designated as distinctly Early Silurian and shallow-marine (Zeigler 1965 [sic]).
>
> The siltstones of the upper member of the Ross Brook Formation occur both as relatively thin beds that are laterally continuous in outcrop and as thicker beds that commonly are lenticular and have uneven lower and upper surfaces. Many of the thicker beds occur sandwiched between layers of thinly interbedded shale and siltstone (Figs. 14 and 21 [sic]). Bottom-surface irregularities characterize many of the siltstone beds and may have resulted partly from erosion by relatively high-energy (storm-generated?) currents.
>
> Some of the siltstone beds occur in groups that form thickening-upward sequences (Figs. 14 and 17 [sic]). Such rhythms possibly indicate seasonal cycles of sedimentation. It is possible that the thicker siltstone beds were

deposited from currents generated by seasonal storms. Groups of beds in which mudstone predominates may denote seasons during which storms tended to be infrequent, or during which the discharge of mud from river-estuaries along the coast was relatively high.

The siltstone beds have many of the characteristics of the storm-related strata described by Lane (Recent Shoreline to Shelf Sedimentary Facies ...., this volume [sic]). The lower contacts of the siltstone beds generally are sharp and the under-surfaces either are flat and structureless or are irregular due to the presence of loaded scour-depressions (Figs. 23 and 26 [sic]). The beds tend to be laminated, with the laminae commonly less than 1 mm thick. Some of the beds lack internal structures, except for parallel-lamination near the tops of the beds where mica content tends to be high. Beds with gently inclined cross-lamination (possible megaripples) are present here and there in the upper member (Figs. 22 and 24 [sic]). The megaripples(?) presumably developed on the seafloor during storms. Composite beds, composed of alternating layers and lenses of parallel-laminated, finely cross-laminated and structureless siltstone, are common (Figs. 16 and 18 [sic]). Many of the beds have sharp upper contacts, but thin gradational zones at the tops of the beds are common as well. The grading is most pronounced in siltstone beds that are mud-rich. Ripple marks and ripple cross-lamination occur at or near the tops of many of these beds and may reflect the waning of storm-generated currents.

These strata contain numerous fossils, which tend to be concentrated at the interfaces of mudstone and siltstone beds. In situ biocoenoses (life assemblages) are found at the upper surface of mudstone beds, whereas transported thanatocoenoses (death assemblages) occur as coquinites at the bases of a number of beds [see Fig. 1 for the location of the first of these beds]. These coquinites consist primarily of carbonate debris (mainly shell fragments) and calcite cement, and typically occur as lenses in scour-depressions (Fig. 22 [sic]).

The mudstones tend to be highly bioturbated and, as a consequence, internal structures are poorly preserved. Vertical burrows are abundant. Suspension-feeding bivalves commonly are preserved in their life positions. The presence of these organisms suggest that relatively clear-water conditions prevailed for much of the time. However, siltstone beds probably were deposited during relatively short-lived episodes of turbulence and high-sediment influx, during which local populations of the bottom-dwelling organisms may have largely perished.

Pickerill (1992) concludes also that the siltstones were deposited by storm-generated currents on a mud-dominated shelf. He believes that the variations in thickness and structures within the siltstone lenses and layers probably reflect variable environmental conditions (storm strength and duration, sediment input, and distance from source and storm). Depth of water could not be determined but he believes the siltstones accumulated above storm wave base above or at fair-weather wave base.

The Arisaig Group is folded into broad southwest plunging folds that are overturned next to the Hollow Fault. The shore section is the southern limb of a large anticline that probably lies offshore to the north. Most argillaceous rock types possess incipient fracture to slaty cleavage that may or may not be axial planar and related to this folding event. In some outcrops near to Arisaig Brook minor open folds with wavelengths of the order 20-50 m and with cross-cutting cleavage can be seen (Fig. 1). Elsewhere bedding-cleavage relationships suggest that the cleavage is axial planar. In other outcrops the bedding and cleavage are kink folded with widths < 5 cm or are crenulated giving a second generation of cleavage.

The Arisaig shore section is generally fractured and cut by several faults with displacements in the order of centimetres to tens of metres. Dip-slip motion is shown on most fault surfaces by slickensides and rotation of fabric suggesting these faults may have resulted from movement on the main Hollow Fault and the uplift of the Antigonish Highlands.

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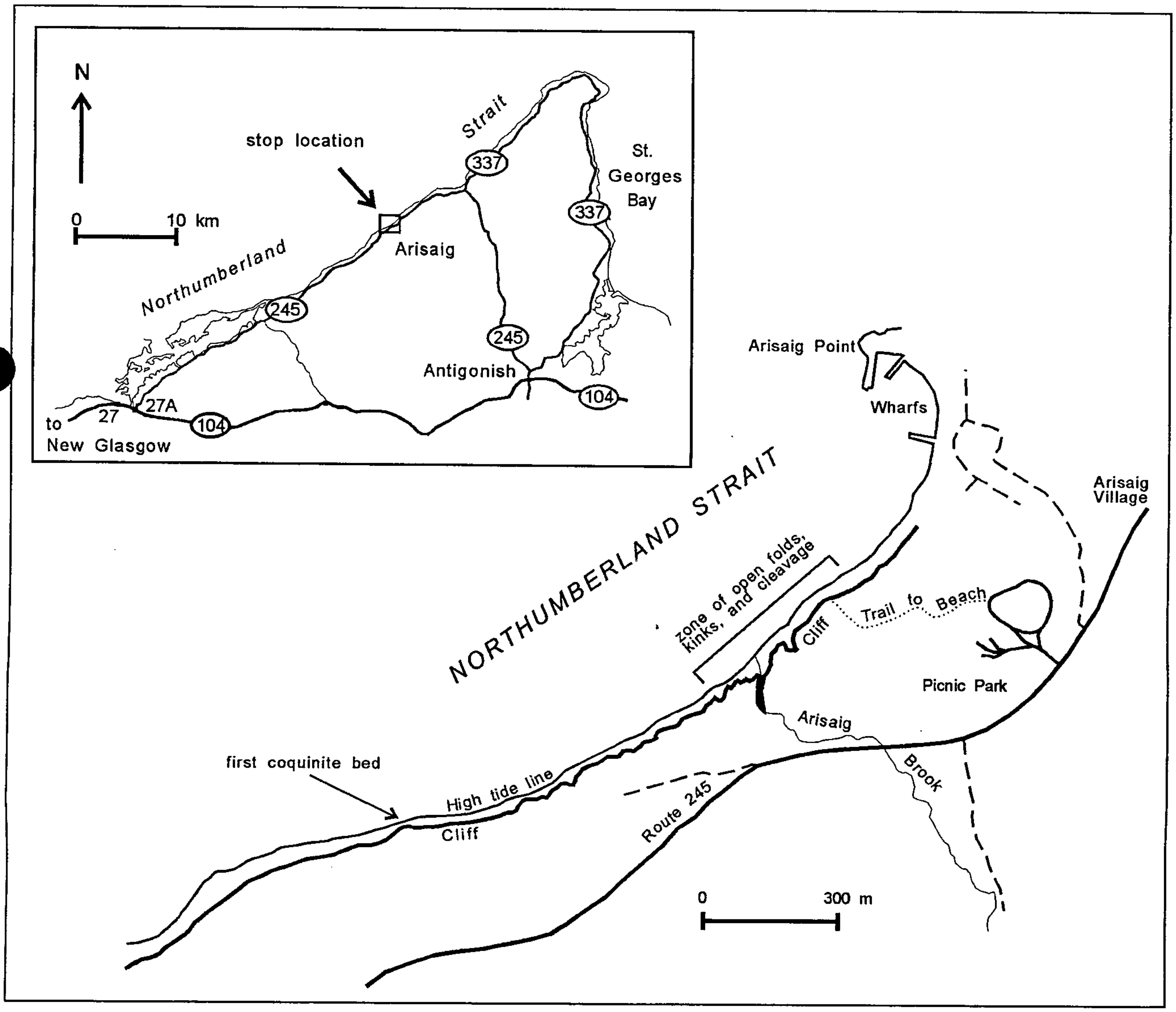

Figure 1: Location map of Arisaig Group sedimentary rocks

Victoria Park, Truro

Understanding the Importance of Rivers

Directions

Victoria Park is located on the south side of the Salmon River Valley in Truro where Lepper Brook tumbles from higher elevations to nearly sea level. It represents a perfect, natural laboratory for the study of streams: both present day Lepper Brook and much more ancient streams.

Highlights

In the upper reaches of the park, the valley of Lepper Brook is a gorge almost 130 feet (40 m) deep. In the lower park, the valley consists of a channel for the brook and a broad, flat floodplain. These features are typical of river valleys and the basis for examining over 360 million years of earth and river history.

Geological History

Victoria Park exposes rocks and land forms that show the effects of erosion and the formation of streams. Two ages of rocks are found in the park. The gorge and waterfalls of the upper park have been created by the erosion of the older, red sandstone and shale of the Horton Group. These rocks belong to the Early Carboniferous period and are about 360 million years old.

The younger of the two groups is the Wolfville Formation. These red-orange sandstones and conglomerates are about 220 million years old and are exposed downstream of the Serpentine Drive bridge. They were deposited in often desert-like conditions during the dawn of dinosaurs in the Middle Triassic period. The two rock units are separated by an unconformity that represents at least 140 million years, when either no rocks were deposited or rocks were deposited but stripped away by erosion.

These two rock groups provide an insight into the continuity of geological processes through time and a marvellous view of ancient stream deposits. They also represent two separate periods of earth history when Nova Scotia was undergoing radical changes.

The story begins 380 million years ago during the collision of two large crustal plates: one was ancient North America and the other was Gondwana (includes Africa, South America, Australia, Antarctica, India and others). The collision produced a range of mountains now called the Appalachians. As soon as the mountains were created, they began to be weathered and eroded. The gravel, sand, and mud coming from the destruction and erosion of the mountains formed layers of sediment deposited by streams which we now call the Horton Group.

After the deposition of the Horton, many kilometres of mud, limestone, gypsum, salt, sandstone, and coal were deposited on top of it. The region was cut and broken by faults, folded, and eroded. Between 180 and 220 million years ago, what is now Nova Scotia lay near the centre of a huge megacontinent (Pangea). Beginning at approximately 225 million years ago, Pangea began to break apart by rifting. As the cracks widened and deepened into valleys, what is now southern Nova Scotia began to split away from northern Nova Scotia. Large amounts of mud, sand, and gravel were deposited in the deep, wide rift valley. Geologists interpret the rocks of the Wolfville Formation to have formed on the slopes of an alluvial fan in desert-like conditions. Lepper Brook exposes deposits of conglomerate (gravel) in the Wolfville Formation that were deposited in an ancient valley cut by Triassic age streams.

Old and New Streams

Streams and their deposits form the continuity between the present and the geological past. Lepper Brook is the present stream in the valley. It probably originated after the end of the last glaciation, approximately 11,000 years ago, as drainage in this area was re-established. But it flows in an valley whose origins date back to the Triassic period.

So, part of the valley where Lepper Brook flows today is quite ancient. How did geologists interpret this from the rocks? The clues are found in both the valley walls and in the streambed of Lepper Brook. In the lower park near the bandshell, the channel of the brook has cut into ancient gravels (now conglomerate) of the Wolfville Formation while the adjacent valley walls expose outcrops of Horton Group shales and sandstones. In order to have younger rock (Wolfville) restricted to a valley and lower in elevation than the older rock, the younger rock must have been deposited in an ancient valley. This would mean that part of the Lepper Brook valley was excavated in the Middle Triassic (about 220 million years ago) and filled with gravels and sands (conglomerates and sandstones) of the Wolfville Formation. After considerable erosion, all that remains are the small exposures of Wolfvllle in the brook near the bandshell.

The Horton Group is the oldest rock unit at 360 million years but it too has abundant evidence of streams and rivers. The sandstones and shales of the Horton Group are interpreted by geologists to represent the deposits left by stream systems over a long period of time. Some of the deposits are thin layers of shale that represent deposits of mud left on the broad flat floodplain. The floodplain is usually muddy and is shown with a 3 on the accompanying diagram; old floodplain deposits of mud are shown with a 7.

Cutting across the flat floodplain is the stream channel which is defined by a curved bottom and usually abundant sand (marked with a 1 on Fig. 2). Older channels are shown at 5. All channels have curved bottoms. Notice how the channels are stacked one on top of another in the diagram. Geologists often find that stream channels will be confined to a certain area.

In many stream systems, such as the Mississippi River, the water flow in the channel is confined by natural levees. When these are breached during major floods, large amounts of water carrying sand and mud flow out of the channel and onto the flat floodplain. When flood waters recede the sand deposit marks the break (see 4). Ancient levee breaks are represented in older rocks of the diagram by flat, parallel sided sandstones layers (see 6).

Descriptions of Stops

STOP 1. Lower Park, Lepper Brook flows out of the narrow, steep-walled valley into a broad flood plain where most of the park's recreational facilities are located. It is possible to distinguish the valley walls, flat floodplain and channel of the brook.

Orange-red coloured sandstone and conglomerate of the Wolfville Formation are exposed in the brook but they may be hidden by layers of sand and mud. The flat-lying layers rest unconformably on steeply tilted sandstones of the Horton Group.

Upstream, the sandstone beds under the bridge provide geologists with a great deal of information. Most of the beds are very uneven in thickness, have curved bottoms and represent shallow water channels of small streams. The finer grained shales in between the sandstone layers were deposited when the river periodically overflowed its banks and covered its floodplain. Gradually, the mud settled, leaving layers of silt and mud.

STOP 2. Pumphouse Exposure. At this location, Lepper Brook valley becomes much narrower, although the brook still has a well-defined channel and floodplain.

The rocks of the Horton Group consist mainly of red-brown shales. The layers are straight and extend from one side of the outcrop to the other. This contrasts sharply with the layers of sandstone under the bridge which pinch and swell in thickness. A few resistant layers of sandstone stand out on the cliff face. These represent flood stage deposits when the river breached its banks and overflowed onto the floodplain. The large amounts or shale indicate that most of these rocks were all deposited on a floodplain.

Notice how much loose rock (talus) has accumulated at the base of the cliff. This rock was broken from the bedrock by freezing and thawing, wetting and drying, and the wedging action of plant roots.

STOP 3. Holy Well Section. This large cliff face offers an excellent view of thick layers of redbrown shale alternating with more resistant, thinner sandstone beds. The shale layers show that the floodplain was occasionally covered by muddy water. Whenever a bank or levee was breached during a major flood, large amounts of fast flowing water and sand poured out of thechannel onto the floodplain. This gave rise to the long thin layers of sandstone.

One of the shale beds contains several "veins" of sandstone cutting across the shale layers. The vein was probably formed when a large crack opened and was filled in by sand. Some geologists believe the cracks were created by the ground motion from earthquakes.

Perhaps you have noticed that the layers of rock are not horizontal, but tilted. This deformation, which resulted in tilting, is the evidence of movement on nearby faults that affected the rocks about 320 million years ago. As the faults moved they created zones of greater pressure within the rock that formed closely spaced fractures called cleavage. Most of the shales here and elsewhere in the park exhibit a cleavage.

STOP 4. Wishing Well Stairs. On the west side of Lepper Brook, a long staircase descends from the hilltop to the brook in a narrow ravine. The brook follows the deep ravine because the action of the water has eroded a soft layer of shale. Running water will often seek the easiest rock unit to erode in order to establish its channel.

STOP 5. Joe Howe Falls. These spectacular falls were created when Lepper Brook could not erode the thick, hard sandstones—it was easier to cascade and fall over these more resistant beds. As you walked up the valley, you may have noticed many resistant layers of sandstone crossing the brook. These deposits of sandstone represent either breaches of the levee (see diagram, number 6) or small channels. Some channels are quite small: only 0.5 m deep and 4 m wide. But the thick sandstones at the base of the falls represent a huge channel 3 m deep and 10 to 15 m wide. This ancient channel could carry more than I 00 times the water in Lepper Brook. Almost all of the rocks located at and above the falls are ancient river channel deposits of sandstone all stacked one on top of another. The placement of the falls at this point was undoubtedly influenced by the large number of vertical cracks in the rock called joints or fractures.

Contributor

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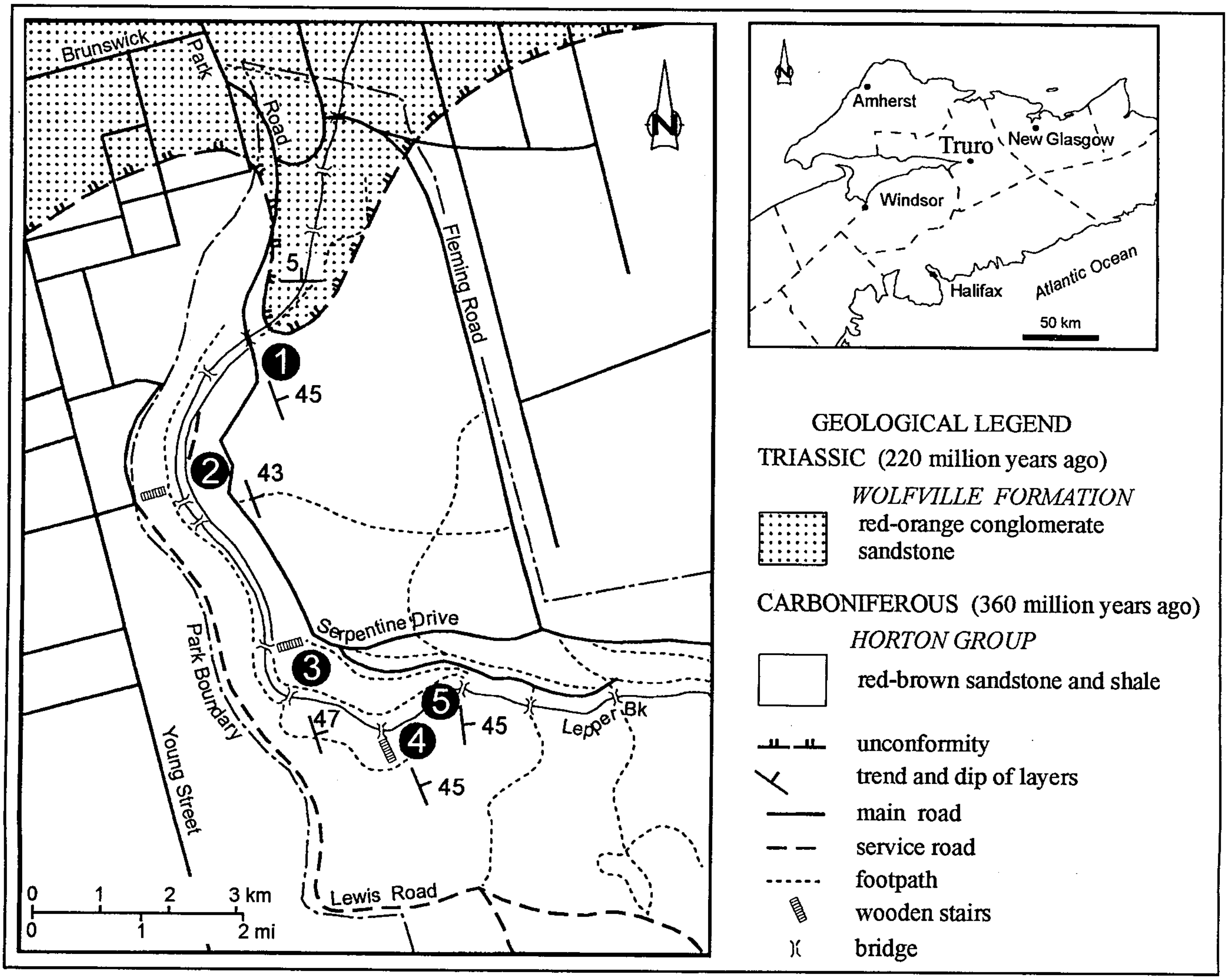

Figure 1: Location map for Victoria Park, Truro, showing bedrock geology and stop locations.

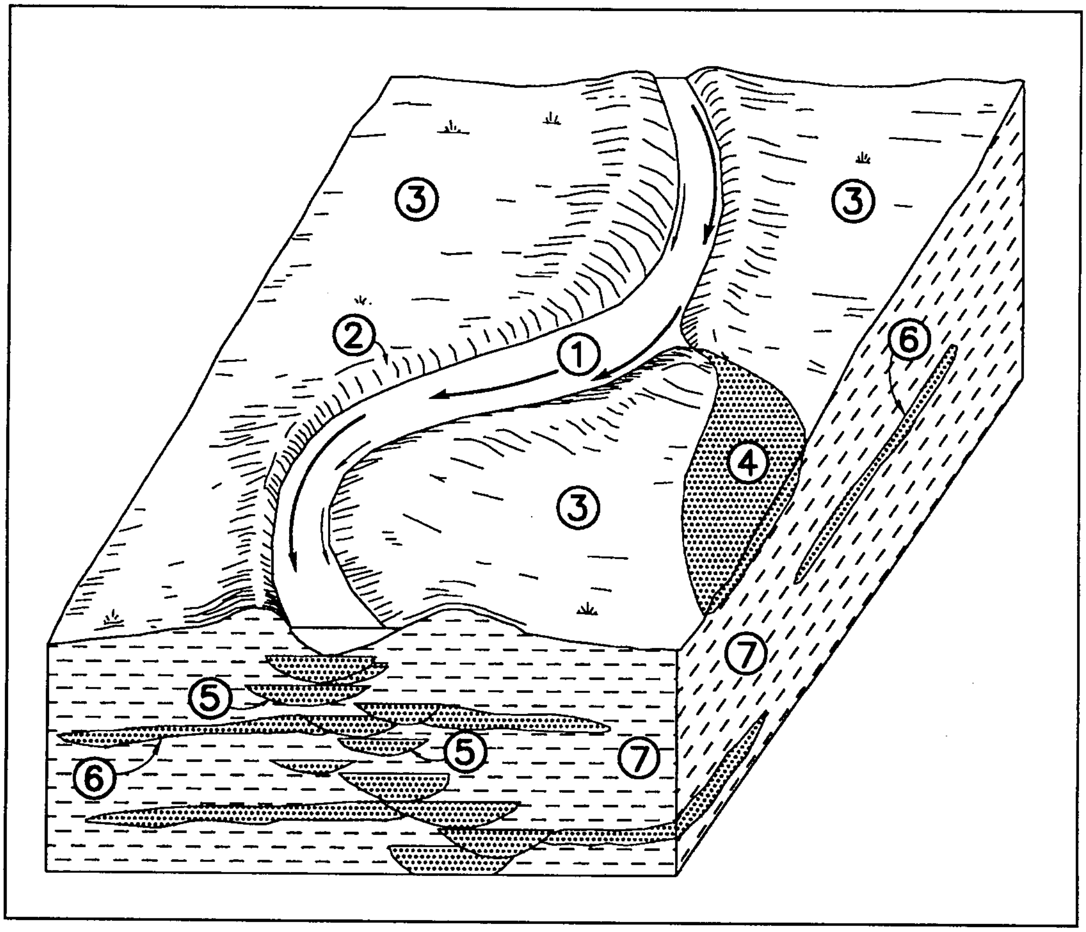

Figure 2: Depositional model of a floodplain showing the different types of deposits.

Five Islands Provincial Park

Triassic-Jurassic Sedimentary and Volcanic Rocks

What's of interest

- raised beach conglomerate (Pleistocene)
- fish-kill bed (McCoy Brook Formation)
- shallow-water delta (McCoy Brook Formation)
- Triassic-Jurassic boundary (Blomidon and North Mountain Fms)
- sand flat-playa/lacustrine cycles (Blomidon)
- domino-style, syndepositional faulting (Blomidon Formation)
- Eolian dune field (Wolfville Formation)

Special Precautions

This trip takes place in an area of extreme high tides (averages 17 metres) and steep spectacular cliffs. Make sure you know the state of the tides at all times and do not get caught behind headlands when it is rising. Follow the tide out. Watch out for falling rocks; stay back from the cliff to make most observations.

Directions

Take Route 2 west towards Parrsboro (Fig. 1), passing Carr's Brook and then up to the top of Economy Mountain, a horst composed of the Fundy Group strata. Stop at the pull-over (see Fig.1) to view the Cobequid Highlands to the north bounded by the east-west trending Cobequid fault (the sharp escarpment in the distance). The valley floor is underlain by the Jurassic McCoy Brook Formation in the foreground, and in the background sediments of the Carboniferous Parrsboro Formation rest against the foot of the fault scarp. Resume driving down the hill almost to the bottom, and turn south into the Five Islands Provincial Park. (Alternately coming from Parrsboro on Route 2 turn into the park after passing by the village of Five Islands.) Follow the park road past the park buildings and park in the lot at the picnic area near the beach. Walk down to the beach, turn left and head southeast (Fig. 2).

Description

The description starts with the youngest rocks and goes down-section, beginning at the Hettangian aged McCoy Brook Formation, through the faulted North Mountain Basalt, followed by the thinned but complete Blomidon Formation, and ending in the eolian sediments at the top of the Carnian aged Wolfville Formation. These rocks are overlain unconformably at the beginning and end of the section by raised beach conglomerates of Pleistocene age and undifferentiated glacial till.

The first outcrops encountered are reddish-brown marginal-lacustrine mudstones and sandflat sandstones of the McCoy Brook Formation (Location 1 on Fig. 2) with few desiccation features. The section is better exposed down shore where it is series of prominent, variably thick, alternating channel sandstones and overbank mudstones (Location 2). The channel sand deposits are characterised by climbing, straight-crested and sinuous-crested ripples, cross- and flaser bedding, truncated foreset beds, and rip-up mud clasts. At the base of these deposits, an intraformational conglomerate composed of these mud clasts also contains fish bones and related coprolitic material. The interbedded, thin overbank/levee silty mudstones are horizontally laminated with occasional widely spaced desiccation cracks, root tubules and burrows. Diagnostic Jurassic-aged vertebrate footprints are also present in the mudstones.

A modern (1992) talus slope has transported to the shoreline large boulders of material from higher up in the channel/overbank sequence (Location 4). The boulders are sometimes composed of a red, cross-laminated sandstone overlying a violet-red laminated claystone. At this contact, whitish, densely packed, complete fish fossils (*Semionotus* sp.) are common and it would appear that this interval is representative of a single fish-kill event.

A shallow-water delta composed of three large-scale stacked channel complexes is encountered at Location 5. Each fining upward sequence of sandstone and siltstone displays reasonably well-developed foresets, though topset beds are usually absent. Evidence for coal and/or other plant material having once been present is lacking.

Northeasterly-trending, down-to-the-west normal faults juxtapose the McCoy Brook strata against the stratigraphically-lower basalts of the North Mountain Formation (Location 6). The basalt breccias exhibit near surface cataclastism resulting from vertical faulting and shearing soon after deposition. Similar faults further west produced the prominent, craggy basaltic stack known as the Old Wife (Location 7). Stevens (1980, 1987) interpreted these basalts and the faults as portions of a tectonically-emplaced dike. Olsen (1989) argues that the absence of chill margins, variably-rotated basalt columns, and chlorite-fibre slickensides infer that the rubble zones are near-surface products of the graben-bounding, left-lateral fault zones which in turn are branches of the larger sinistral Portapique Fault a few hundred metres north of this site. These are the basalts in which excellent samples of zeolites such as stilbite, chabazite, analcime, heulandite and gmelenite associated with magnetite and quartz (amethyst) can be found in cracks and vugs along the outcrop, and amongst the boulders of recent rock falls. Looking back to the northwest from this point, a photogenic view is presented which well displays the stacked geometry of the deltaic channels in about +35 metres of McCoy Brook strata.

Turning eastward, gently north-dipping palya-lacustrine sediments of the underlying latest Triassic Blomidon Formation are seen in fault contact with the basalts. The Triassic-Jurassic boundary is beautifully exposed in the cliffs to the east, where the dark brown basalts overly the bright orange-red Blomidon sediments. The 1–2 metre thick light green-grey banded unit at the very top of the sediments actually represents this boundary, and contains probable charred organic material (Location 8).

Mertz and Hubert (1990) identified and measured over 100, 1-2 metre thick sand flat sandstone/ playa-lacustrine sandy mudstone-claystone cycles in the Blomidon Formation (Location 9). These cycles can be laterally traced along the entire 2 km of exposed section here. The sandstones tend to be the more resistant beds, reflecting increased evaporite mineralization, mostly gypsum. The trend of these cycles over time in reflecting wetter climatic conditions is revealed in an increase in the thickness, frequency, and total volume of red lacustrine claystones, along with a corresponding decrease in the above parameters for the sand flat sandstones and the playa mud flat sandy mudstones.

Walking eastward and downsection, the large Blomidon blocks and boulders on the beach permit detailed examination of the bed relationships and the characteristic sedimentary features in the lithologies of each cycle component. Sandpatch and crumb fabrics are particularly well displayed in this setting (Location 10).

Nearing the end of the Blomidon outcrop sequence, alternating beds of orange sand flat sandstone and red lacustrine claystones/playa sandy mudstones reflect their lower position in the section, in that they are dominated by the former lithology (Location 11). Domino-style antithetic faulting is clearly exposed in these sediments immediately adjacent to, and soling into, a normal fault which separates them from the stratigraphically-lower Wolfville Formation eolian sandstones (Location 12). This site displays beautifully the intimate and dynamic relationship between sedimentation and tectonics. Individual beds within the fault slivers can be traced laterally into undisturbed strata and uninterrupted sedimentation eventually onlapped and buried the faulted sediments.

Spectacular exposures of Wolfville Formation eolian dunes can be seen extending for several kilometres to the east (Location 13). The total thickness of these sandstones is estimated at 33 metres and is stratigraphically at the top of the formation. The maximum thickness of an individual cross-bedded set is about 3 metres, averaging 1.6 metres. Minor fluvial sandstones beds are also present. First-order erosional surfaces, which cross-cut all other eolian features (second- and third-order surfaces), generally have a relief of 5-20 cm and in some cases display ventifacts and deflation-lag surfaces. Second-order surfaces bound wedge-and/or tabular-planar, and tangential cross-bed sets which reflect changes in the paleo-wind direction and the resultant alteration in the direction of dune migration. Dune avalanche slip faces are representative third-order surfaces. It is believed that this sequence represents an accumulation of compound transverse-type dunes with a paleowind blowing from the northeast.

The paleodrainage patterns recorded elsewhere around the Minas Basin by Hubert and Forlenza (1988) led them to conclude that the drainage in this area was to the east, towards Truro (Fig. 3) whereas in the Wolfville and Cape Blomidon regions the drainage was to the west. This places the drainage divide somewhere in the vicinity of the present day Avon River estuary and along a line oriented to the north. The rock types recorded in the sediments are indicative of a semi-arid environment although localized and earlier wetter conditions existed to allow coarse grins and fauna to be found lower in the sequence and to the south (see Fig. 3). These wetter and humid conditions are also seen in the Scots Bay sediment, above the basalts.

Sinistral transverse motion along the major east-west Cobequid fault resulted in considerable relief with micro-grabens/basins to the south. Discharge from seasonal rainfalls in the uplands transported to the basins large particles in a matrix of sand and silt where they were deposited in a series of coalescing alluvial fans. Eolian deposition was probably due in part to the persistence of north-northeasterly derived trade winds which parallelled the strike of the pre-Triassic massif. These winds would have reworked the material from the fans and plains as they dried up. In summary, the rapid vertical changes in depositional environments from stops 1 to 13 reflect a larger picture of a variable climate and landscape during Triassic-Jurassic time.

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Source

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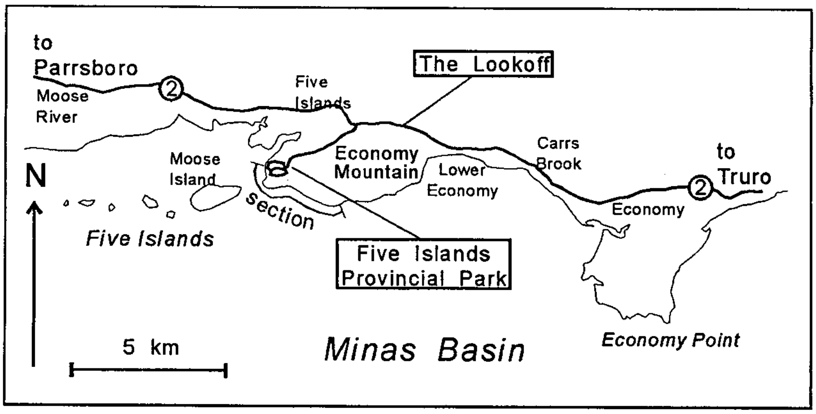

Figure 1: Location map of Five Islands Provincial Park.

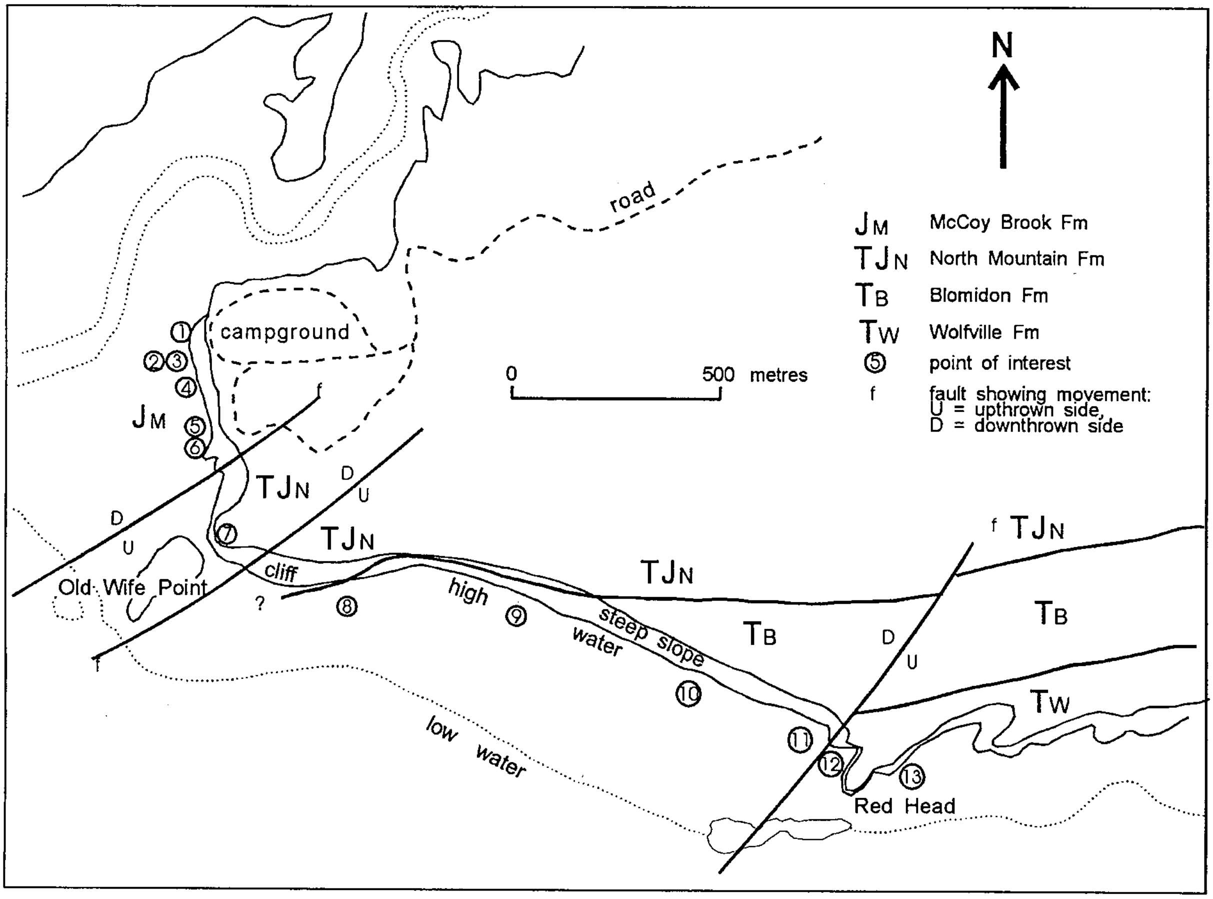

Figure 2: Detailed geology map of the Five Islands Provincial Park area.
From Brown and Grantham, 1982

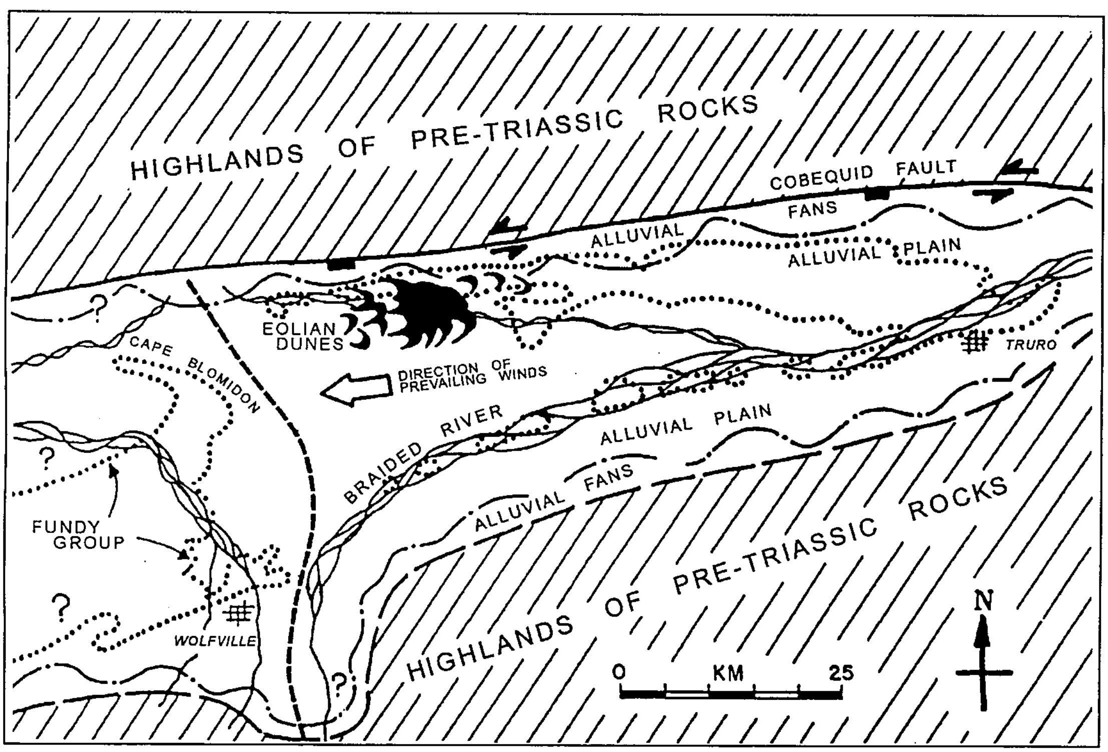

Figure 3: Paleogeography map of the Minas Basin area during deposition of the Wolfville Formation in late Triassic time. From Brown and Grantham, 1992

Spicer Cove and Squally Point

Volcanic Rocks, Dikes, and Thrust Faults

Location and access

This field trip is to the coastal section between Spicer Cove and Squally Point. Access is from Highway 209. At Squally Point, there are magnificent outcrops of early Carboniferous volcanic rocks of the Fountain Lake Group, cut by dikes and northward-directed thrust faults (Piper et al. 1996a).

About 0.5 km south of Apple River, take the paved road to West Apple River, on the coast of the Apple River estuary (Fig. 1). At about 10 km from Highway 209, the unpaved continuation of this road crosses the mouth of Cove Brook at Spicer Cove and then turns uphill and inland. Walk westwards along the beach from Cove Brook for about 2.2 km to reach Squally Point.

Cautions

Much of the coastline beyond 0.5 km from the mouth of Cove Brook is inaccessible around high tide. First access is about 2 hours after high tide at Digby, but depends on the spring-neap cycle and weather conditions.

Much of the section is in coastal cliffs, with the usual danger of occasional falling rocks. The red sandstone and conglomerate cliffs 0.5 to 1.1 km west of the mouth of Cove Brook are particularly unstable: these rocks should be examined where they are exposed on the foreshore and visitors should avoid the foot of the cliffs, especially in the spring.

General description

Three main rock units are seen on the 2.2 km walk from the mouth of Cove Brook to Squally Point (Fig. 1). For the first 0.5 km, there are low cliffs and foreshore outcrops in the Ragged Reef Formation (Ryan et al. 1990) of Westphalian B age, comprising grey mudstones and sandstones and a few coaly layers. Plant fossils are common in the Ragged Reef Formation.

An easterly trending fault separates the Ragged Reef Formation from an unnamed red unit of conglomerate and sandstone to the southwest (Salas 1986). This unit is probably equivalent of the Falls Formation in the eastern Cobequid Highlands (Piper et al. 1996b). Walking westward one goes progressively lower in the stratigraphic section. Paleocurrent indicators suggest transport to the northwest. The base of the unit rests unconformably on the Fountain Lake Group volcanic rocks at Squally Point. Clasts in the basal part of the conglomerate include cataclastic and mylonitic granite, both felsic and mafic volcanic, rocks, and grey quartz wackes. None of the thrusting found in the Fountain Lake Group at Squally Point has affected this unit. This unit thus appears to post-date the early Carboniferous deformation of the Cape Chignecto pluton (Waldron et al., 1989, Koukouvelas et al., 1996) and the Fountain Lake Group.

The volcanic rocks of the Fountain Lake Group at Squally Point, described by Piper et al. (1996a), consist of generally rhyolite shallow intrusions, minor flows, and pyroclastic rocks interbedded with a few basalt and andesite flows (Fig. 3). Fault-scarp breccia, fluvial conglomerate and sandstone, and lacustrine siltstone and limestone are interbedded with the volcanic rocks. The volcanic sequence is cut by flat-lying, north-vergent, thrust faults and high-angle N-S wrench faults, both of which have acted as pathways for emplacement of diabase dikes. The oldest dikes predate the thrusting. The volcanic pile is hydrothermally altered, some of the mafic racks are in places highly silicified and some rhyolites show ochrous weathering of sulphides. One rhyolite near Squally Point has yielded a U/Pb age on zircon of ~353Ma (Greg Dunning, pers. comm. 1996). The Fountain Lake Group is overthrust by granite of the Cape Chignecto pluton at Seal Cove, 3 km south of Squally Point (Piper et al. 1996a). The Cape Chignecto pluton has been dated at 361± Ma (Doig et al. 1996). Chemical composition of the volcanic rocks is similar to that of intrusive rocks of the Cape Chignecto pluton.

Principal outcrops of interest (keyed to Fig. 1)

1. Grey sandstones and siltstones, with plant fossils, and rare coaly layers, Ragged Reef Formation.
2. E-W trending fault separating the Ragged Reef Formation from the Falls Formation equivalent. When there is not too much sediment on the foreshore, barite is visible along this fault.
3. 10 to 50 m high cliff section in Falls Formation equivalent. Avoid this unstable cliff and examine the outcrops on the foreshore. Sandstones and conglomerates were deposited in an alluvial fan or fluvial environment.
4. Unconformity of Falls Formation equivalent over altered and fractured rhyolite of the Fountain Lake

Group. Unconformity dips about 30° SE and is repeated by a small NE-trending fault. (Tbe eastern outcrop of the unconformity is partially obscured by a rock fall, demonstrating the need to avoid this unstable cliff).

5. A three-metre thick succession of siltstone, conglomerate and breccia, with vertical bedding, is preserved along a wrench fault separating porphyrific rhyolite in the east from flow banded rhyolite in the west. A diabase dike, with chilled margins, has intruded between the base of the sediment succession and the flow banded rhyolite (Fig. 2). This dike forms a small headland.

6. A prominent N-S diabase dike, forming a headland, is cut by a flat-lying thrust, visible as you walk towards the headland from the east.

7. Immediately west of (6), amygdaloidal basalt exposed on the foreshore overlies the autobrecciated top of a rhyolite flow. The top of this basalt, exposed in the low cliff, is intensely silicified, and is overlain westward by a prominent red rhyolite.

8. Keyhole arch. The rock here is a fragmental lithic tuff, including xenoliths of cherty siltstone, probably a Precambrian basement lithology.

9. Immediately west of Squally Point, numerous diabase dikes cut the rhyolite sequence, which is principally a welded tuff or ignimbrite. The geometry of the dikes shows evidence of substantial E-W extension (see Fig. 6 of Piper et al. 1993). Silicified basalt is exposed on the foreshore.

10. At Squally Point, note the raised beach at the top of the cliff, dating from the early post-glacial highstand of sea level, when the land was still depressed as a result of ice loading. This raised beach is illustrated in the AGS video "The Last Ice-Age."

If time and tide permit, walking 0.5 km south of Squally Point to Anderson Cove will take you through more exposed volcanic rocks.

11. The rocks on the foreshore west of Squally Point expose rhyolite interbedded with red mudstone and cut by diabase dikes.

12. Andesite flow with well-developed pillows, suggesting extrusion into a local lake. [loc. 29]

13. Green hyaloclastitic rhyolite, in places with lithophysae, formed by extrusion of rhyolite into a local lake. [loc 27]

14. A 4 m-thick, poorly sorted conglomerate, interpreted as a lahar (volcanic mudflow) overlies flow-banded rhyolite to the south.

15. South of the brook about 0.5 km south of Squally Point, porphyritic rhyolite is cut by a prominent thrust plane at the base of the cliffs, which ramps up northward through the cliff immechately south of the brook. This thrust ramp has been occupied by later diabase dikes (Fig. 3).

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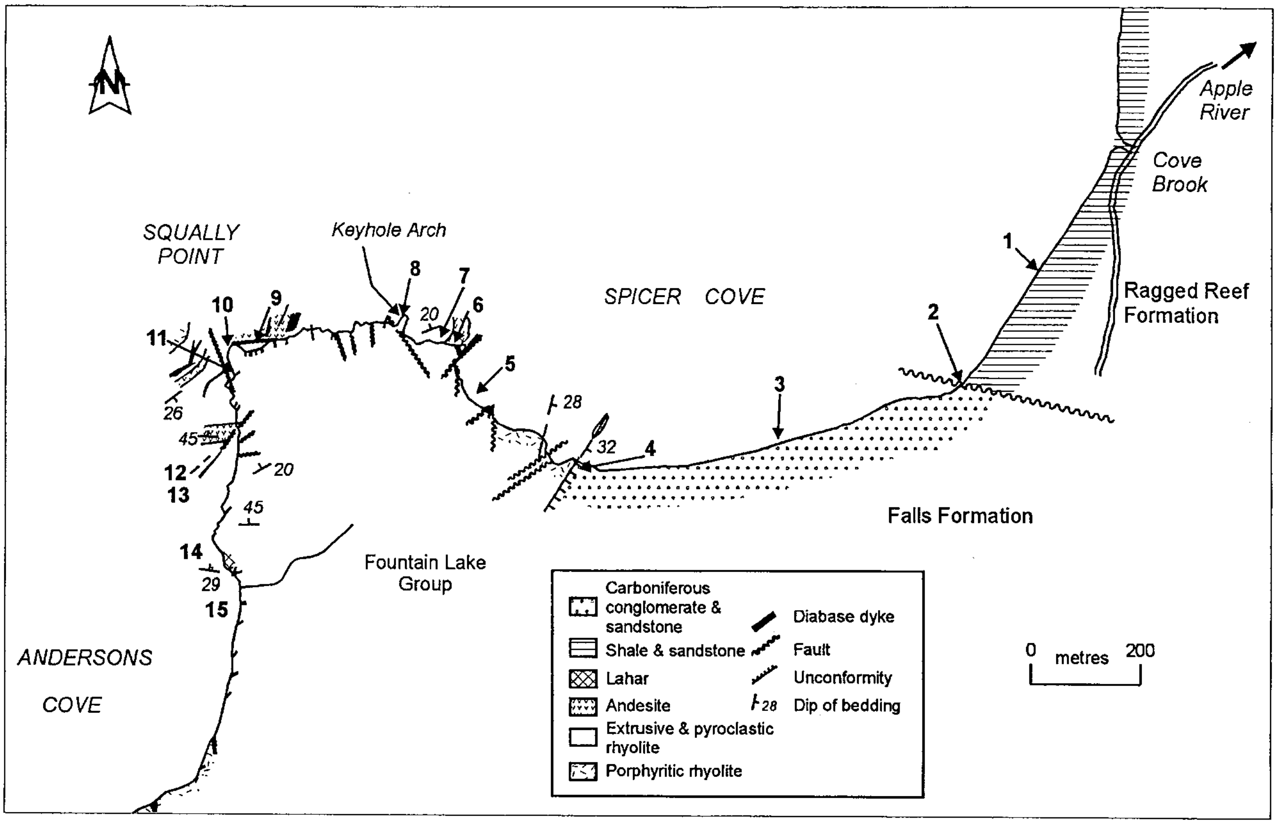
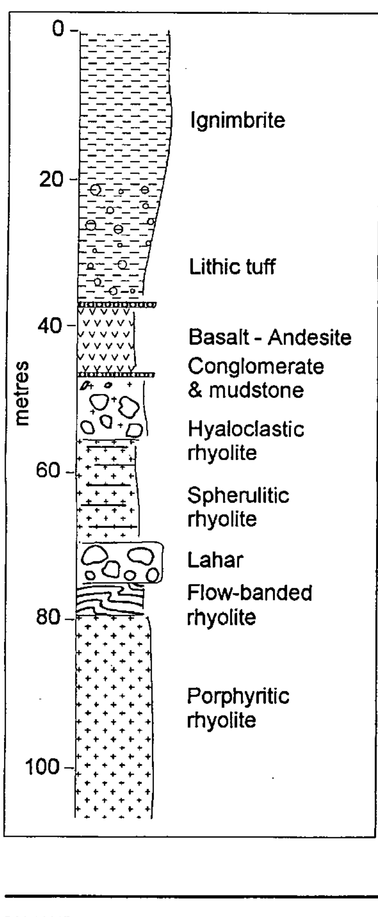


Figure 1: Map showing access to the Spicer Cove–Squally Point section from Apple River



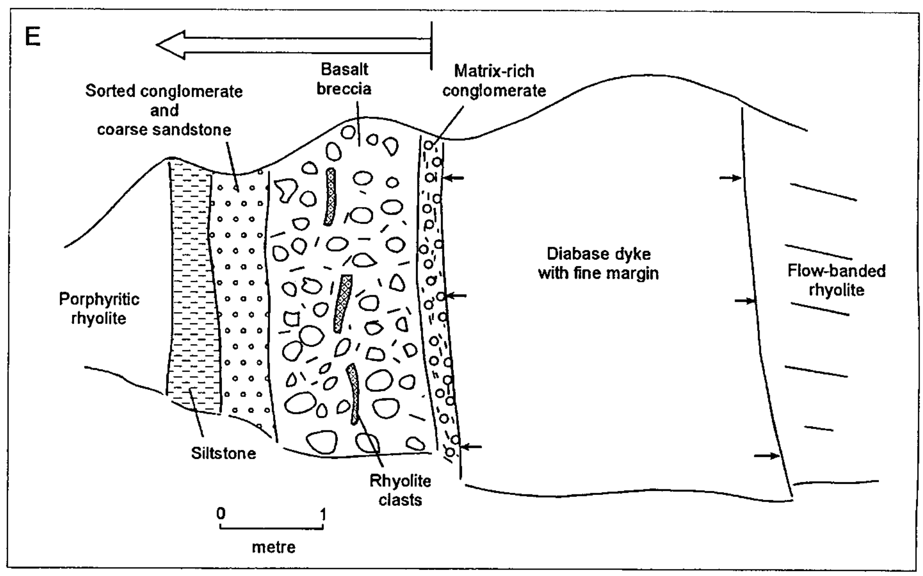


Figure 2 (left): Generalised stratigraphic section of the volcanic rocks near Squally Point (Fig. 7 of Piper et al., 1996a).
Figure 3 (above): Sketch of field relationships seen at outcrop 5 (Fig.9 of Piper et al., 1996a).

Joggins

The Fossil Cliffs of Joggins—One of the World's Classic Geological Sites

Location

The Fossil Cliffs of Joggins loom above the Bay of Fundy on the western coastline of Cumberland County. This majestic coastal exposure of Carboniferous age rocks stretches 50 km from Boss Point north of Joggins, to Spicer Cove and Squally Point in the south, near Cape Chignecto. It is the part of the section between Lower Cove and Ragged Reef that is commonly referred to as the Fossil Cliffs (Fig. 1). The village of Joggins lends the cliffs their name and is the geographic location. Joggins is situated in western Cumberland County, northern Nova Scotia, on Route 242. From Amherst, travel south on Highway 2 and turn right onto Route 302 at Upper Nappan. Proceed south to Maccan, turning west (right) onto Route 242. From Springhill, travel north on Highway 2, turning west (left) at Little Forks and then north on Route 302 through Maccan. At Maccan, turn west (left) on Route 242 to Joggins. From Parrsboro, travel north on Highway 2, keeping left at Newville (Halfway) Lake over the Boar's Back Road to River Hebert. At River Hebert, proceed west (left) on Route 242 to Joggins.

Historical Significance

The fossil cliffs of Joggins were brought to the attention of the scientific world through the remarkable Nova Scotian physician-cum-geologist, Abraham Gesner, in the early 1800s. The detailed description of the cliffs, a painstaking layer-by-layer description, was the very first field project of the newly formed Geological Survey of Canada and its new director, Sir William Logan, in 1843. Logan's account still stands as the reference description for the great sequence of sedimentary beds exposed in the cliffs. A monument commemorating this first project at Joggins, erected 150 years later by the GSC, can be seen in front of the Joggins Fossil Centre on Main Street.

It was the great passion for the cliffs by a young Nova Scotian that would establish the place of Joggins in the world's great geological sites. William Dawson, who would one day be knighted for his remarkable contributions to paleontology and geology, studied the cliffs over a period of four decades. In the process, Dawson made many discoveries of great scientific importance, many of which stand today. Dawson's mentor was the most influential geologist of the time, Sir Charles Lyell, of Britain. Together, Lyell and Dawson, in 1852, made one of the most astonishing paleontological discoveries of the day, and one that continues to captivate the imagination. Inside the stone-filled interior of an upright and once hollow fossil tree, the two geologists discovered the bones of a land-dwelling vertebrate. Prior to this, it was believed that animals had only adapted to life on land much later in geological time. But there was more: the skeleton was that of a reptile, and for almost a century and a half it remained the oldest known reptile ever to have been found on earth; only recently was an older find made at East Kirkton, Scotland.

What has continued to intrigue paleontologists and naturalists is the manner in which these skeletons came to be inside the once hollow trees. The most popular theory and one suggested by Dawson, is that the animals fell into the hollow trees once the surrounding sand and mud had built up to the rim of the hollow tree stump. This "pitfall" theory has been widely accepted and pictured in fossil books and scientific papers alike. A new look at the specimens collected by Dawson and Lyell, now in the collections of the Natural History Museum, London, and at the Redpath Museum, Montreal, together with research of the fossil forests in the field, has revealed some startling new insights. Nearly all the tetrapods are found in the base of the trees, within a mixture of limy silt and fossil charcoal. Were the animals living inside a tree hollowed by fire, or perhaps were they overcome by smoke from a prehistoric forest fire? Whatever the answer, it is clear that other animal life made the hollow trees their home, including armoured millipedes and the oldest known land snails, both discovered by Dawson. In all likelihood, so too did the tetrapods.

Paleontology

The fossil record at Joggins is nothing short of remarkable. The most prolific fossil evidence comes from the wetland forests. From great standing lepidodendrid trees with their long, pockmarked Stigmarian roots to the fern-like foliage of seed fern trees, black coalified plant fossils abound. Most are found as plant litter within sandstone, and represent stems and branches that collected at the bottom of stream beds. The different lepidodendrid trees can be distinguished by the patterns of their fossil bark: *Sigillaria* (parallel, vertical ribs with alternating leaf scars); *Lepidodendron* (elongated diamond-shaped markings); *Lepidophloios* (shorter diamond shapes) as well as *Paralycopodites,* identifiable in part by large branch scars or "knots" on its stem.

The black beds of coal, often stained yellow by the high sulphur content of the Joggins seams, are themselves fossils: fossil peat, the waterlogged organic soil that under the right conditions builds up beneath swamps and bogs.

Within the black, limy beds that commonly overlie the coal seams, the shells of mussel-like bivalves occur in abundance. These beds, known locally as "clam coals," formed at the bottom of organic rich lakes or estuaries. A profusion of aquatic fossils have been found within these clam coals, including fishes, sharks, rays, and amphibians. Barbed spines of primitive xenacanth sharks and their whitish cone-shaped coprolites (fossil excrement) often can be found; in these beds along with the pelecypods and countless tiny, oval shells of ostracods (water fleas). Shiny brown, diamond-shaped scales of fossil paleoniscid fishes and large (loonie-size) roundish scales of crossopterygian fish lie scattered on the bedding surfaces. The disarticulated scattered remains of the aquatic life is testimony to the voracious scavenging that must have taken place on the bottoms of these bodies of water.

Much more rare is the fossil record of animal life that flourished in the wetland forests. Bones are rarely found outside the fossil trees, but there have been some spectacular, articulated skeletons of anthracosaurs, an evolutionary intermediary between amphibians and reptiles, discovered in recent years. Individual footprints of these and other tetrapods (four legged vertebrates, including both amphibians and reptiles) range in size from mere millimetres to 16 cm or more. The discovery of any bone is potentially of great scientific importance, and the finder should report such finds to the Nova Scotia Museum, Halifax, or to the Fundy Geological Museum, Parrsboro. The staff of the Joggins Fossil Centre can offer helpful advice in this regard.

Rarest of all is the fossil record of the huge winged insects that whirred through the forest primeval. A rare example is the dragon fly-like *Megasecoptera* in the collection of the Joggins Fossil Centre on Main Street. Most insects of the Carboniferous bored and sucked the nutritious spores and seeds of the Coal Age trees with a dagger-like proboscis and were themselves the prey of agile needle-toothed reptiles.

Fig. 2 represents the column of rocks that make up the Joggins Section, with the coal seams numbered by Sir William Logan as bookmarks. Beside the column are symbols that represent the various discoveries of fossil animal life over the past 150 years. Inside the column, the position of upright fossil trees is depicted: at least 48 different fossil forest horizons have been identified from Lower Cove to Ragged Reef.

Site Description

The layers of sedimentary rocks that form the Joggins Cliffs from Lower Cove to Ragged Reef are tilted, or dip southward at about 20° from the horizontal. Each bed was deposited after the one lying underneath, and so the age of the rocks becomes younger as we proceed south. The total thickness of beds of rock from Lower Cove to Ragged Reef is 2100 m. The great thickness of sediment that formed these rocks was deposited over a period of about 2 million years. This represents one of the very thickest accumulations of sedimentary rocks known over this span of time. The age of the rocks from Lower Cove to about 100 m north of Dennis Point on the north side of MacCarron's Creek is late Westphalian A (after the Westphalia coal area of Germany), whereas those south of the point are of the next youngest age, early Westphalian B.

At this period of Earth History, approximately 310 million years ago, Joggins lay near the equator, in one of several intermontane basins that stretched across southern New Brunswick, northern Nova Scotia, and Prince Edward Island. Long rivers coursed through the interconnected valleys on their way to a distant sea, thought to have lain somewhere to the northeast of the Maritimes. The Cobequid highlands, to the south of Joggins and to the west of the Caledonias, had risen by Carboniferous times to mountainous heights and partly bounded the lowland area to the northeast, known as the Cumberland Basin. Mountain runoff deposited gravelly fans, now conglomerate, at the foot of these mountains. The sedimentary rocks of the Cumberland Basin underlie much of northern Cumberland and Colchester counties, and include the Springhill and Joggins-Chignecto coalfields.

If you begin your journey at Lower Cove and proceed south to Bell Brook, you will follow the depositional history of the beds. The colour of the cliffs often appears reddish brown due to the wash of glacial till and soil above. The actual colour of the rocks is best seen where they have been washed clean, as in the intertidal zone or at the foot of the cliffs after a particularly high tide. The colour of the rocks tells a tale. Grey shale and mudstone beds were continuously waterlogged muds, with the darker grey rocks generally indicating a higher content of organic material. Black beds such as coal are highly organic-rich. Reddish brown beds (red beds) on the other hand reflect the fact that the sediment on occasion dried out after it was deposited. Red beds therefore can record drier, more seasonal climate periods, whereas the grey beds can reflect humid periods during which the mud remained wet.

The beds of the Joggins Section have long been thought to have been deposited by continental rivers and lakes, with no record of marine conditions. Certainly, this is generally true, because no deep or 'open' marine fossils have been found. However, recent research has shown that the manner in which some rocks were deposited may reflect the response of rivers to changes in the ancient sea level somewhere beyond the Cumberland Basin, and there is new fossil evidence to suggest that some of the beds were deposited in coastal estuaries (see Field Stop 5.2, the Forty Brine Paleovalley). Indeed, some of the fossils from Joggins are strikingly similar to life forms that live today in coastal waters.

Selected Field Stops

The sheer dimension of the Joggins Section is overwhelming. The following portions of the section are suggested as points of interest on which to focus in order to help understand the section as a whole.

Fundy Forests
A good example of two of the major plant types of the Carboniferous and the conditions under which they grew.

One of the best exposures of fossil forests is midway between Lower Cove and Bell's Brook. The location is easily found nearby a waterfall that issues from old mine workings on the Fundy seam, the iron-rich waters staining the cliff bright orange. Old wooden pit props and iron rails protrude from the cliff at three levels of abandoned underground mine-workings, bearing witness to the forces of erosion. In the strata immediately below, several upright fossil trees occur, mainly along two horizons 9 m apart.

The trees are lycopsids, the typical Coal Age trees also referred to as lepidodendrids. It is difficult to identify the large trees at the bases of their trunks, which is usually all that is preserved, but most of the Fundy forest trees probably were of the genus *Sigillaria*. The large lycopsid trees had shallow root systems that were adapted to waterlogged soils. They were quickly suffocated, however, by the blankets of mud and sand deposited by flash floods.

Protruding from the thin sandstone beds around the erect lycopsid trees are the jointed, bamboo-like stems of *Calamites*. These smaller trees thrived in such areas because of the ability of their branches, if buried, to sprout "adventitious" roots and to continue growing upward through the sediment. Such growth patterns of the calamites, evocative of "snakes and ladders," are visible in the thin sandstone beds through which the erect lycopsids protrude. The great lycopsid trees, once buried, would have died, whilst the hardy calamites would have continued to grow.

Forty Brine Paleovalley
Example of response of the landscape and sediments to probable sea level change.

The Forty Brine Seam (Coal 20 of Logan's section) is one of the most widespread coal seams in the Cumberland Basin. It is characterized by a thick resistant limestone roof (overlying bed). This limestone bed has been one of the most productive sources of aquatic fossils at Joggins (Fig. 2). The 80 cm-thick seam is located stratigraphically below (to the north of) the first thick multi-storey sandstone body south of Lower Cove and the Fundy Forests (Fig. 1). The roof of the Forty Brine Seam is an 80 cm-thick, dark grey limestone that bears abundant fossils of mussel-like pelecypods, tiny ostracods and fish remains; a similar bed lies approx. 6 m above, and forms a prominent, resistant black ridge that stretches across the intertidal zone of the shore: look for it as a marker. Between this bed and the trace fossil-bearing sandstone described next, is a sequence of grey mudrocks that are conspicuous only by their lack of plant fossils. A thin, flat bedded sandstone that lies exposed in the cliff 30 m above the Forty Brine coal seam thickens to form a small headland and reef on the foreshore. This flaggy sandstone bed has yielded the trace fossils (trails) of horseshoe crab-like limulids and other invertebrate trails that have been interpreted as the signature of an ancient estuary. Within the grey mudrocks above this bed, roots and bark rinds of lycopsid trees are common. Looming above these beds is a very thick multi-storey sandstone bed in turn overlain by reddish mudrocks.

The section is interpreted as a near-coastal peatland (Forty Brine Seam) drowned by a rise of water levels, perhaps related to rising sea level and the transgression of the sea (the limestone roof). The limestone roof and pelecypod-bearing clam coal above it represent basin-wide events, the nature of which remains unclear, that, led to temporary cessation of peat accumulation. The faunal record of these clam coals (see Fig. 2) includes paleoniscid (bony) fish, sharks, and even rays, which traditionally have been thought to have lived in fresh, rather than sea water. Deepening water is suggested by the absence of plant fossils in the mudrocks above. The trace fossil bed may represent estuarine sand bars and the highest level reached by the sea. The resumption of growth by lycopsid trees above the trace fossil bed suggests that sea, level dropped, exposing the muddy sediment to colonizing plants. As the sea withdrew, a major sandy river (thick channel sandstone body) carved a valley downward through the sediment, and the reddened mudrocks formed on the better drained landscape. Tetrapod footprints have been found in these red beds, testifying to their having once been the ancient surface of the land.

Coal Mine Point—Site of the tetrapod-bearing trees of Dawson and Lyell.

The first and most prominent rocky headland to the north of the Bells Brook entrance is Coal Mine or Hardscrabble Point. Here, a very thick multi-storey sandstone body forms a resistant promitory. *(Beware the slippery rocks.)* The best vantage point for interpreting this part of the section is on the north side of the point. It was here, in 1851, that Sir Charles Lyell, visiting the younger Sir William Dawson, stumbled serendipitously upon the bones of amphibians and the world's earliest reptile-within the cast of a once hollow tree that had tumbled out of the cliff.

The thin coal beds beneath the Coal Mine Point sandstone originated as low-lying, mineral-rich peat swamps. On the surfaces of the coal beds can be seen long, ridged structures of the bark of fallen lycopsid trees that originally formed the peat. Above the coal beds, which are splits of the Kimberley Seam, two resistant sandstone ridges, or reefs, extend from Coal Mine Point into the intertidal zone of the Bay of Fundy. The lower (northern) reef is pierced by countless holes, a few centimetres in diameter, where the casts of calamites have been eroded from the sandstone. These small calamite trees formed an incredibly dense thicket, with many stems per square metre.

Beneath the lower sandstone reef, a fossil forest of lycopsids stands entombed, stretching out into the intertidal zone. From this fossil forest over a century ago, Dawson obtained more than 100 skeletons of tetrapods, representing several species of amphibians and two species of reptiles. From one tree alone, Sir William found 13 individual skeletons! Dawson's explorations of the reef were greatly assisted by the local coal mine manager who supplied dynamite to the cause!

The very thick sandstone body at Coal Mine Point represents a later, major river channel, perhaps formed similarly to the thick channel sandstone above the Forty Brine Seam. Some of the large fallen blocks of sandstone at the base of the cliff are impressed with large, twin tracks (looking very like those made by a bulldozer) believed to have been made by the gigantic myriapod *Arthropleura,* crawling at river's edge. Coaly stems and trunks of trees within the sandstone recall trees swept into the river.

Cautionary Notes

Tides. The tides of Fundy are renowned for their range, and all trips to Joggins must take tide times into consideration. As a general rule of thumb, allow yourself two to four hours to pass from Lower Cove to Bell's Brook. There is nowhere in between that one can escape the rising tide. If you are somewhere midway along the cliffs, you should leave no later than 2 hours before high tide to return safely. In the event of an emergency, do not attempt to scale the cliffs! A fall could well be fatal. Proceed to the nearest rocky point and climb atop the large blocks and wait out the tide. Better to be late than never to return home.

Falling rocks. Avoid standing at the foot of the cliffs. Small rocks rain down continually, and from a height of one hundred feet, even a small rock could cause serious, perhaps fatal, injury. Windy days are often the worst for falling stones. Hard hats are recommended. Rarely, but inevitably, large falls of rock from the cliffs thunder down without warning, and even a hard hat will not save you from such a rockfall. It is best to examine at close range the rocks on the shore and the long 'reefs" that stretch out into the intertidal zone.

Slippery footing. The rocks underfoot can be very slippery, especially when the weather is damp. The rocks that are covered by green algae are especially slippery!

Collecting Restrictions

The Fossil Cliffs of Joggins have been designated a Special Place by the Nova Scotia Museum, and there are certain restrictions to collecting fossils. Most importantly, you are not allowed to remove any fossil from the cliffs or reefs (in other words, no hammering at the cliffs) without a Heritage Research Permit, which can be obtained from the Curator of Special Places at the Nova Scotia Museum, Halifax. Technically, any fossil is the property of the people of Nova Scotia, but the Nova Scotia Museum allows the casual collection of common fossils. Fossils that might be scientifically important should be brought to the attention of the Museum, so that the world can share in the knowledge that might be gained from them.

Suggesting Reading

General reading

The Fossil Cliffs of Joggins, by Laing Ferguson, 1988. Nova Scotia Museum, 52 p.

Dawning of the Dinosaurs, by Harry Thurston, 1994. Nimbus Publishing and Nova Scotia Museum, 91 p.

One of the Greatest Treasures ... The Geology and History of Coal in Nova Scotia, by John Calder, Kevin Gillis, Don MacNeil, Rob Naylor and Nicole Watkins-Campbell, 1993. Nova Scotia Department of Natural Resources Information Circular No. 25, 42 p.

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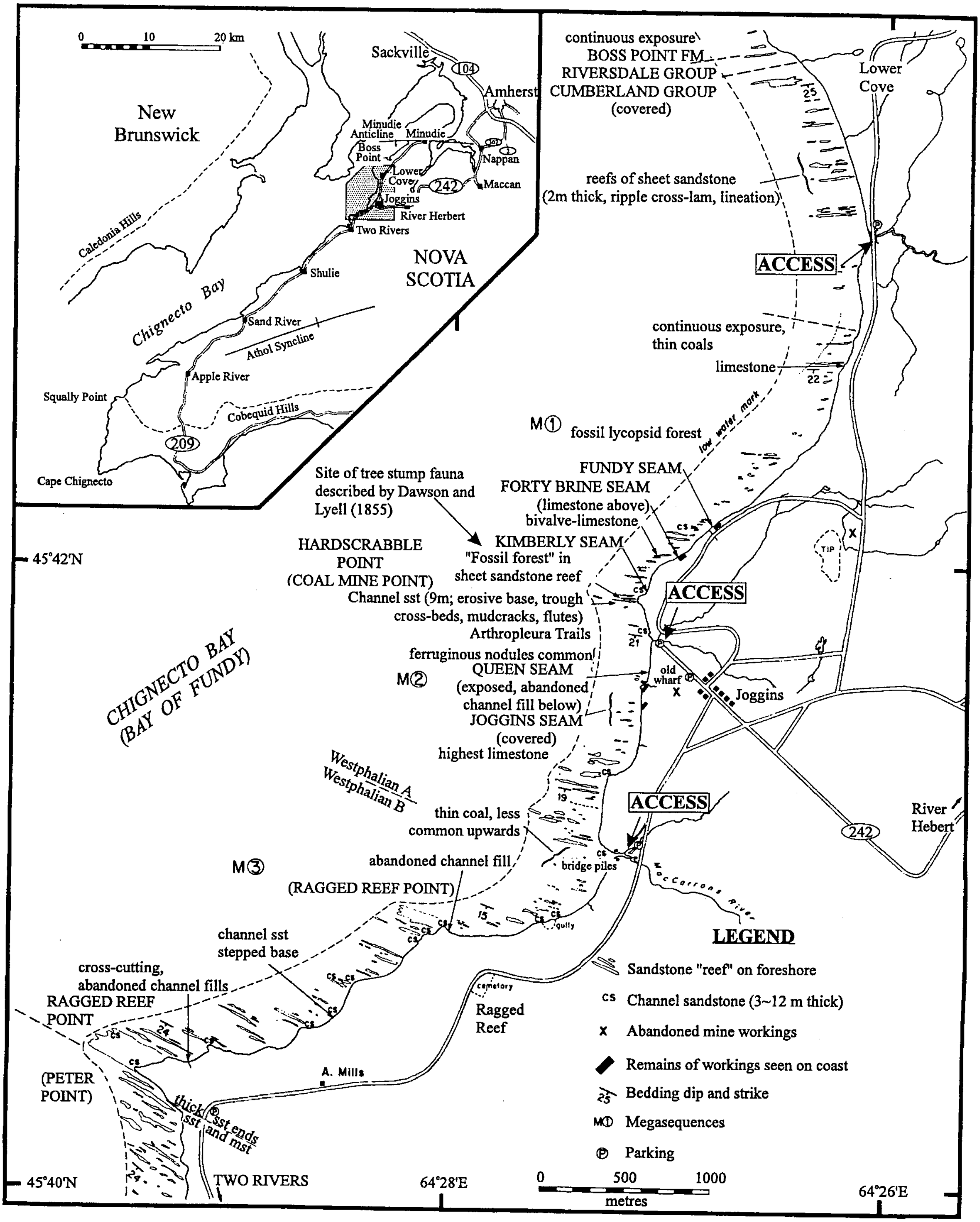

Figure 1: Map of Joggins area showing geology and access routes.

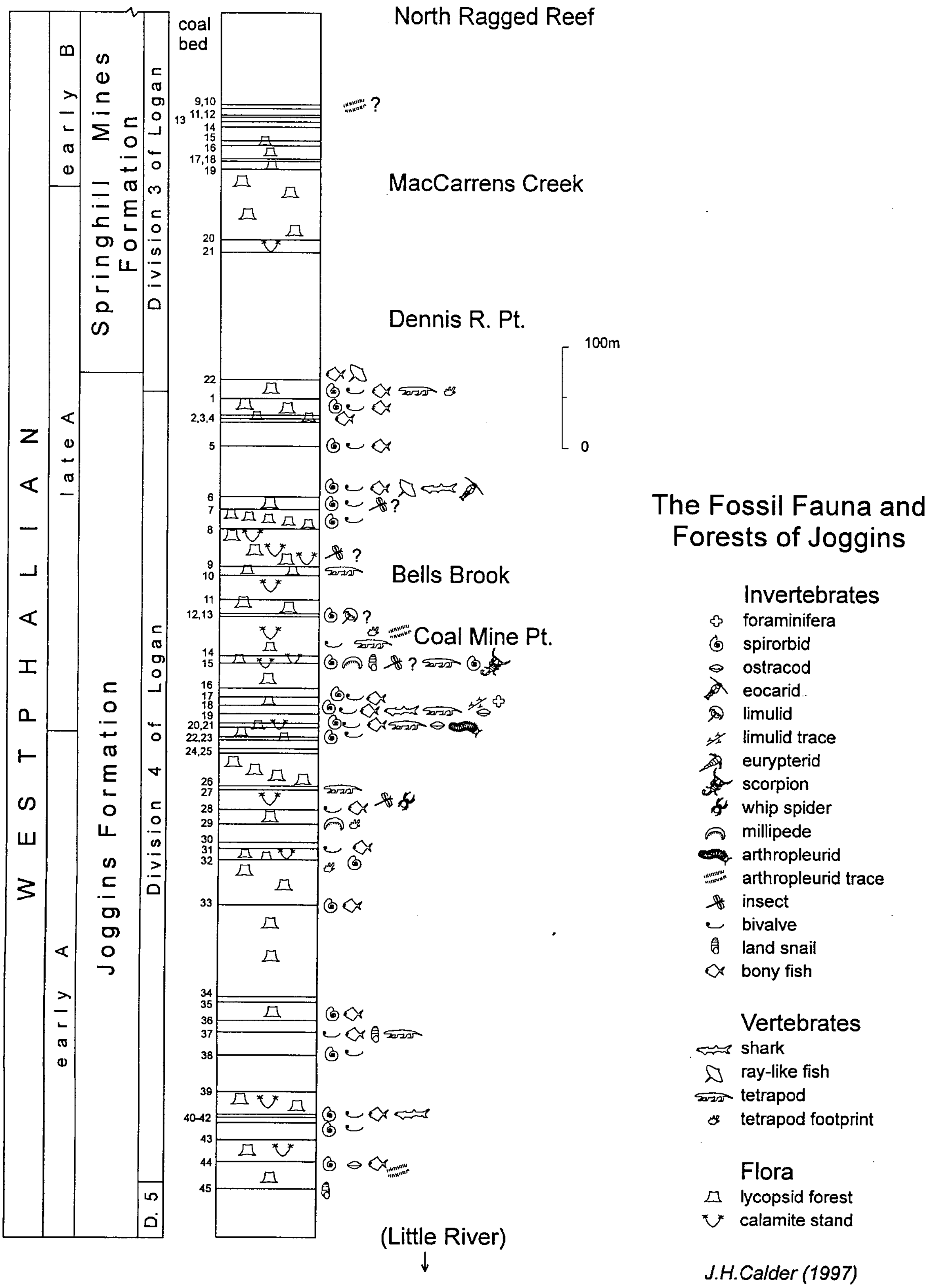

Figure 2: Stratigraphic column of the Joggins Section, the coal bed numbers are by Sir William Logan, the fossil data is compiled from the past 150 years.

Rainy Cove, Pembroke

The Carboniferous-Triassic Unconformity

Location

Rainy Cove is located on Fig. 1 at Pembroke on the coast of the Minas Basin. Travel from Halifax on Highway 101 to exit 4. Turn east on Highway 1 approximately 4.6 km to Newport Corner where Route 215 starts. Turn left onto Route 215. Pass through Brooklyn (5.4 km from the start of Route 215) and continue northward crossing the Kennetcook River at 19.4 km) and the community of Cheverie at 34.7 km. Continue to 45.0 km where Route 215 crosses a small stream (Rainy Cove Brook). Turn left onto the dirt track immediately east of the brook. There is currently (1995) parking for three or four cars at the end of the track adjacent to the beach.

Walk north along the beach, swinging east around the first headland. This headland may be impassible for up to two hours on either side of high tide; it is best to time your visit for a falling tide. Unstable cliffs are a second hazard at this location. Do not approach closer to the base of a cliff than the height of the cliff. Numerous boulders weighing many tons can be found on the beach; most of those you can see have fallen from the cliffs during the last ten years!

General Description

Rocks of the Horton and Fundy groups are exposed along the coast of the Minas Basin between Cheverie and Walton, showing spectacular structural relationships. The Horton Group is of Early Carboniferous age (Tournaisian). It consists mainly of sandstones and shales with minor conglomerates, deposited in a variety of non-marine, mainly lacustrine environments. Horton Group rocks are among the first units that convincingly span the boundary between the Meguma and Avalon Terranes of the Appalachians, showing that by Early Carboniferous times the two terranes had collided and were experiencing similar geologic histories. However, deformation of rocks along the boundary continued throughout the Carboniferous Period. The effects of some of this deformation are spectacularly seen on the shore of Minas Basin.

Younger rocks of the Fundy Group are also present at Rainy Cove. These are of Triassic age. They record an episode of extension that immediately preceded opening of the Atlantic Ocean. A major fault formed along the old terrane boundary; the crust immediately south of the fault was down-dropped, forming the Fundy basin. In this basin accumulated the Fundy Group, consisting of mainly red conglomerates, sandstones, and shales, together with thick basalt lava flow that now form prominent cliffs at Cape Blomidon. Along the south shore of the Minas Basin, the southern edge of the Fundy Group, which is relatively undeformed, can be seen resting on the highly deformed Horton Group rocks, forming a relationship known as an angular unconformity.

The rocks exposed in the cliff (Fig. 2) belong to the Early Carboniferous Horton Group. They consist of sandstones and shales with rare dolomitic carbonate-rich beds. These strata represent sands and muds deposited in non-marine environments—probably in shallow lakes and on lake shorelines. If you are lucky you may find fossils of plants that grew on the banks or drifted into the lake and were buried with the sand and mud. Vertical beds approximately 50 m beyond the end of the path contain numerous oval fossil tree stump fossils, and a conspicuously mud-cracked horizon.

The Horton Group at this location is principally noteworthy for its deformation: the strata were tightly folded during deformation which probably took place later in the Carboniferous period. As you walk north, you can see numerous tight folds and associated faults.

The last major fold is a cliff-high antiform—a fold shaped like an upturned V. This fold has a plunge—its hinge line slopes out of the cliff downward towards the beach. If the tide is reasonably low you will be able to see the outline of the fold traced in ridges of resistant sandstone that cross the shore. There are places to the north and south of the fold where it is possible to deduce the younging direction—the original way up—of the strata from cross-laminations and other structures preserved in sandstone beds. These show that the strata to the north, though near vertical, are actually slightly overturned. The younging direction is actually towards the core of the large antiform. Thus the beds in the crest of the antiform are actually completely upside-down, and the youngest rocks are found in the core of the fold, making it a syncline. This is an example of what is known as a downward-facing fold: one in which the rocks in the core are overturned. It shows that deformation was more intense here even than is apparent at first glance; the beds had probably already been turned upside down by a much larger structure before the fold that you see in the cliff was formed.

North of the large fold, the strata continue in a near-vertical orientation. However, at the top of the cliff a new unit of sedimentary strata can be seen, dipping gently to the north. These strata belong to the Fundy Group, of late Triassic age, about 150 million years younger than the Horton Group. They were deposited on the upturned edges of the Horton Group strata which had been subjected to previous erosion. This type of structure is known as an angular unconformity; this is one of the best exposed angular uconformities to be seen anywhere. This same surface is exposed at several other locations along the Minas Basin shore; it can be seen at Whale Cove to the east, for example, and near Walton. Do not get too close to the unstable cliffs to view the unconformity. With a little searching it is usually possible to find an outcrop of the contact on the rocky part of the beach well away from the cliff.

Notice that the Horton Group rocks immediately below the unconformity surface are reddened and weathered by erosion that occurred in the Triassic Period. The overlying Fundy Group strata are mostly conglomerates, with minor sandstones. Adjacent to the contact the conglomerates contain mainly angular fragments of the immediately underlying Horton Group. Further up-section, the conglomerates are more rounded and show better sorting. They are strongly cross-bedded; the outlines of dunes and trough cross-beds may be seen in the cliffs at the northeast corner of the cove and in the adjacent foreshore. These conglomerates were deposited in either a braided river or an alluvial fan environment.

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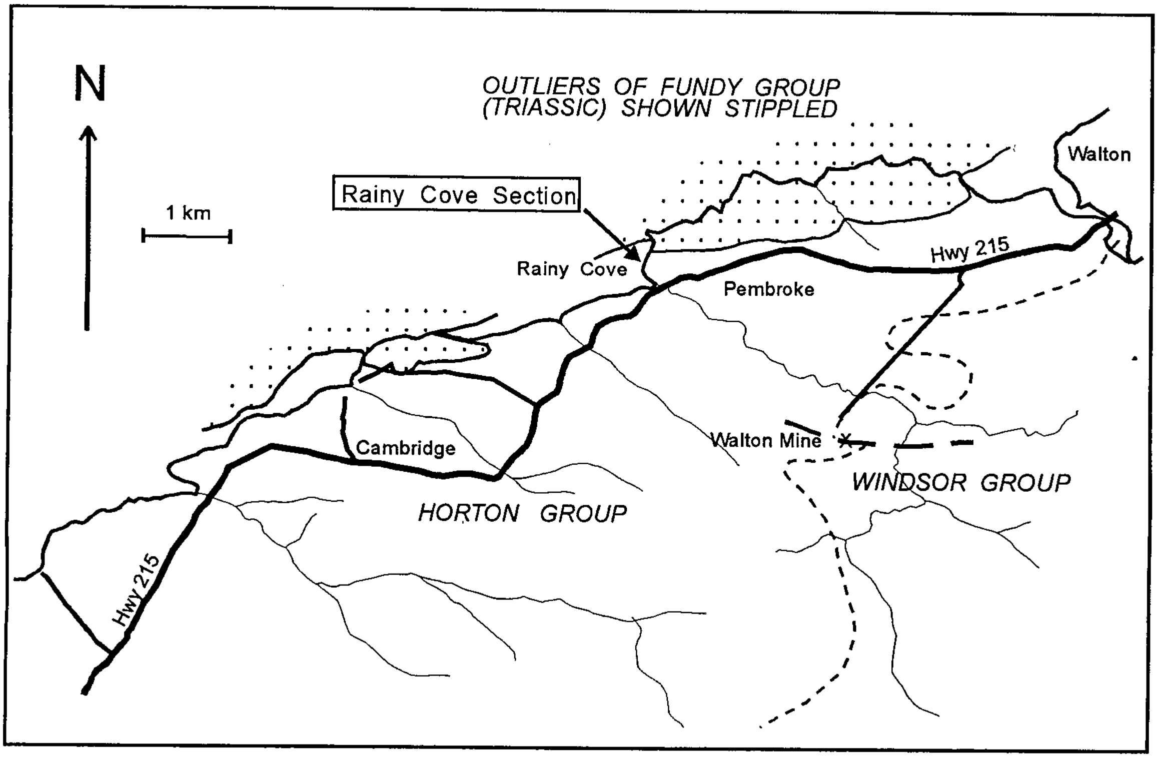

Figure 1: Location map of Rainy Cove with outline of general geology of area

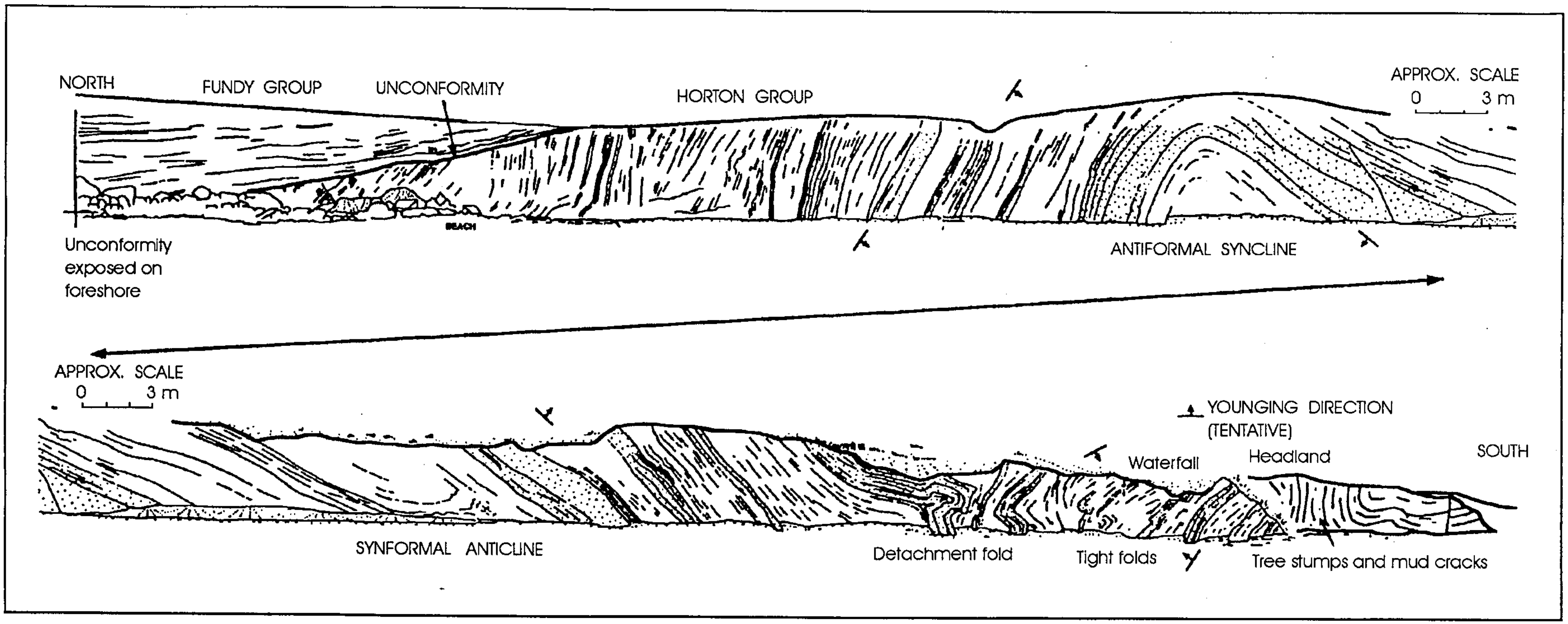

Figure 2: Detailed geology of the cliff at Rainy Cove. Traced from photographs taken in 1995

Horton Bluff—Blue Beach

An Ancient Lake or Restricted Marine Bay

Purpose

- To observe ancient sedimentary rocks.
- To observe fossilized remains of organisms and plants.

Directions

Horton Bluff-Blue Beach is on the western side of the Avon River estuary, north of Hantsport, Nova Scotia (Fig. 1). To get there take exit 9 at Avonport, off Highway 101. Turn left at the stop (first T-junction) and then about 300 m along take the right at the next T-junction onto Bluff Road. Follow this road around the bend past the school and then over the railway tracks. About 50 m past the tracks there is a short dirt road going to the shore (called the wharf road, but it is not posted); this is the northern access to the bluff. You can park here on the dirt road taking care not to block access and walk down the shore to the rocks.

Alternately you can continue 2 km further to Blue Beach Road. This is a dirt road that goes east off Bluff Road for about 800 m, under some railway tracks and then down to the beach. Park in the area on the other side of the underpass, making sure you do not block a driveway. Driving further down to the beach is not recommended in any circumstances. At the beach, walk north to view the rocks, exiting at the wharf road and walking back along the paved road or back along the beach, the way you came.

Precautions

If you wish to see most features, plan on a whole day's outing but beware that even if you are not planning on looking at the whole beach, this is not an easy walk. The area has extreme high tides (17 m at Hantsport) and steep spectacular cliffs. There is no other easy access once you have left the road areas so make sure you know the state of the tides at all times. Follow the tide out and do not get caught behind headlands when it is rising. Watch for falling rocks at all times; stay back from the cliff to make most observations. The best rocks are found on the intertidal areas which are slippery when wet and muddy at best so wear appropriate footware. The intertidal area is exposed to winds and the sun: remember to take lots of water and a lunch, wear appropriate clothing. Practise extreme caution at all times.

Introduction

The Horton Bluff Group is part of a large stratigraphic sequence in Atlantic Canada that covers an area about 1000 by 400 kms. In the Minas Basin it has a pronounced unconformity on Lower Palaeozoic rocks of the Meguma Terrane and is in turn conformably to unconformably overlain by the Early Carboniferous Windsor Group. The Horton Bluff Group is the basal rocks of post-Acadian Orogeny time (mid-Devonian to early Carboniferous) that filled several extensional or half-graben basins formed by faulting along the Cobequid-Chedabucto fault and associated systems. These basins were quickly filled with sediments and this sedimentary tectonic process continued up into the early Permian.

The rocks exposed at Horton Bluff-Blue Beach are the sedimentary rocks of the Horton Bluff Formation which formed in the early part of the Carboniferous Period. The rocks contain a variety of well preserved sedimentary structures, fragmentary plant and animal fossils, and trace fossils (amphibian tracks, root traces). These were all deposited in a large, wave-dominated lake and/or restricted marine bay that was gradually infilled by deltaic and fluvial sediments from the south-southeast.

William Logan, one of the earliest members of the Geological Survey of Canada, visited the site in 1841 and found some of the oldest amphibian tracks (*Hylopus logani*) in North America (the oldest are found in Greenland). The prints are small, about 2 cm wide and slightly longer, with three major toes and a smaller one on the side. The trackway is about 8 cm wide and the stride about 20 cm. Since then numerous other trackways have been found, some up to several metres in length with prints 8 cm long. Many trackways are reminiscent of kick marks made by floating animals.

The rocks of the 2 km long cliff at Horton Bluff–Blue Beach are part of the Horton Bluff Formation and are divided into two members (Fig. 2). The Blue Beach Member is characterized by an abundance of grey clay shale, ostracods, and disarticulated fish material. Dolomitic concretions and beds and the lack of non-stratified mudrocks, thick sandstone units, and sideritic nodules are also important in distinguishing this member from the overlying Hurd Creek Member. The Hurd Creek is generally coarser grained with coarsening upward sandstone- and clay shale/siltstone-dominated cycles. In this area the Blue Beach section is about 160 m thick and the Hurd Creek 100 m.

These rocks are best described by Martel and Gibling (1996) and are what follows. The Blue Beach Member has repeated asymmetric, coarsening-upward cycles, 1–22 m thick (average 6 m) which contain characteristic rock types in a more or less ascending order. The bottom of the cycle consists of dark grey, fossiliferous and fissile clay shale with minor siltstone lenses which grades up into siltstone and very fine grained sandstone interbedded with clay shale. Subaerial and shallow water features such as mud cracks, rain drop imprints, and symmetric ripples, are common and generally increase upward. Lensoid, scour-based bodies of very fine sandstone and siltstone form the lowermost strata and contain hummocky cross-stratification (HCS), planar lamination, and wave ripples. Sole structures, including recurved groove casts and clastic dikes are also common. The mudcracks are filled by fine sand and silt from the top down and look like little daggers in cross-section whereas the dykes were formed by water and gas pressure escaping upwards through cracks in the overlying beds. The dikes are typically folded by later compaction of the surrounding mud/clay and some are quite large, from several cms to nearly a metre long.

Wave-rippled, very fine- to fine-grained sandstone, in bedsets 5 to 20 cm thick, typically overlies the HCS-bearing strata. Planar-bedded siltstone caps most occurrences of these. These beds have yielded agglutinated foraminifera and ostracods with hypersaline and fresh-water affinities. The tops of the cycles consist of severely disrupted, olive grey to green mudstone, in many cases interbedded with tabular-bedded siltstone and/or massive, laterally continuous, tabular dolomite, and nodular dolomite (septarian nodules) within disrupted mudstones. Vertical casts of *in situ* tree trunks associated with subaerial features such as mudcracks are preserved locally within the top beds. Other plant material ranging from spores and rare cones to whole trucks several metres long are seen throughout these beds as well as in those above.

These features in the Blue Beach Member are characteristic of a large, wave-dominated lake that was at least periodically connected to the ocean. The cycles are progradational in style and represent an upward transition from offshore clay shales to nearshore/shoreline sandstones and siltstones to paleosols (ancient soil horizon). The grey clay shales were probably deposited below wave base in oxygenated waters allowing scavengers to survive, but the fissile nature of the shales suggests that interstitial waters were anaerobic. The presence of wave ripples, HCS and recurved groove casts indicates that wave action was prominent in the shallow-waters.

The Hurd Creek Member contains two types of cycle: a coarsening-upward, sandstone-dominated cycle and clay shale/siltstone-dominated cycle. The sandstone-dominated cycles are up to 16 m thick and consist of a basal clay shale overlain by one or a combination of: (A) planar- and lenticular-bedded siltstone and clay shale; (B) flaser-bedded sandstone; and (C) interbedded rippled sandstone and clay shale. These are in turn overlain by coarsening-upward, medium- to very coarse-grained sandstone with planar and trough cross-stratification in sets up to 25 cm thick; the sandstones are up to 8.5 m thick and commonly laterally continuous. The sandstones are generally well-sorted, quartzose, with varying proportions of intraclasts of paleosol nodules, oolitic limestone and mudstone. It is within the Hurd Creek Member that the amphibian trackways have been found. In the northern area of exposure are several large sedimentary collapse structures that appear as circular or domal beds planed off by the later non-effected beds. They range in diameter from 1 to 15 metres, and are thought to have been produced by liquified or gas-charged sediment escaping upward through a vent, resulting in the dome-like structure.

The coarsening-upward clay shale/siltstone cycles resemble the cycles of the Blue Beach Member. They make up a small proportion of the Hurd Creek type section, but are common or even predominant in other sections. The Hurd Creek Member was deposited within a large, wave-influenced standing body of water. The coarsening-upward cycles indicate repeated progradation of coastal sediments over offshore sediments, as for those of the Blue Beach Member. However, the cross-bedded sandstones that cap many cycles are interpreted as shoreline attached, sand bar deposits. Some sandstones with deeply erosional bases and moderate sorting may represent deltaic distributary-channel deposits.

The rocks are fairly flat lying and continuous throughout the whole section except in the area of the lighthouse (see Fig. 2). Here the beds have been kink folded with associated low-angle reverse faults along kink boundaries or separate from the kinks. Many of the faults terminate parallel to bedding and there is ample evidence that the whole section has experienced bedding parallel as well as across bedding micro-faulting. Many of these are now infilled with thin veins of crystalline calcite.

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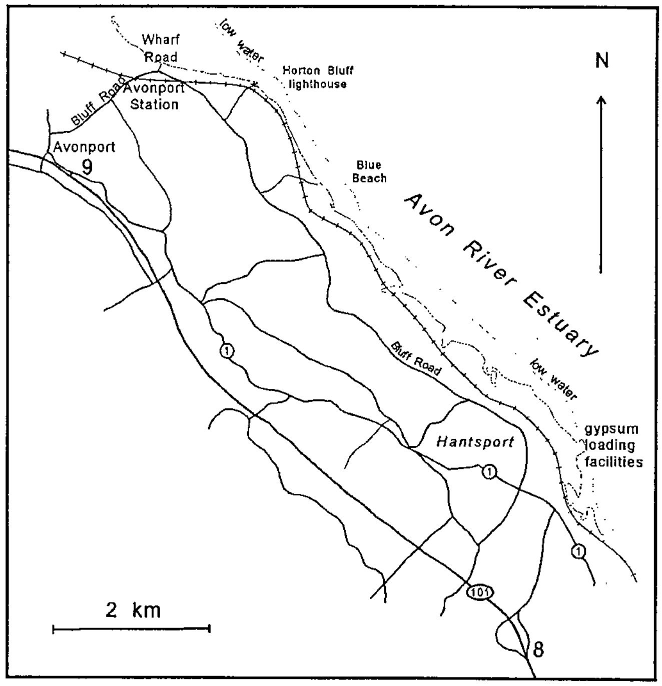

Figure 1: Location map for Horton Bluff to Blue Beach

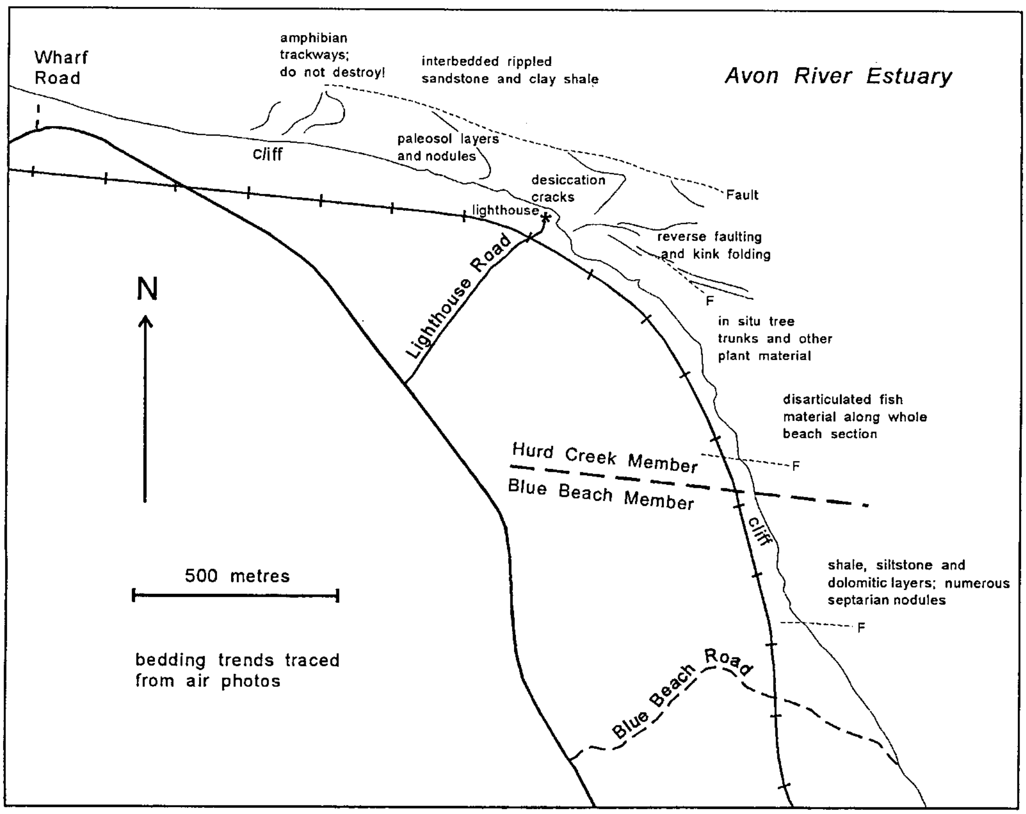

Figure 2: Detailed geological map of Horton Bluff to Blue Beach

Ross Creek

A Section Through a 200 Million Year Old Basaltic Lava Flow

Summary

An easy walk of about 300 m along the shoreline southwest of Ross Creek reveals a section through a 200-million-year-old basaltic lava flow. The lower part of the flow consists of medium-grained basalt with columnar joints, and cut by mineral veins. The upper part of the flow consists of very fine grained and spectacularly amygdaloidal basalt. Limestone with chert nodules overlies the basalt, and was deposited in a lake on top of the basalt flow. Excellent mineral samples, including zeolites, chalcedony, and amethyst, can be collected on the beach from veins, amygdules, and chert nodules weathered out of the outcrops.

General description (Figs. 1 and 2)

Park at the top of the final steep descent to the shore, and walk down the gravel road to the beach where Ross Creek flows onto the shore. For this field trip, you will be walking south from Ross Creek. The time required is about 1–2 hours, depending on how much mineral collecting you do! Note that this section can be examined only at mid- to low tide. At high tide you cannot walk around the cliffs.

Introduction

At the end of Permian time (245 million years ago), the continents were joined in one large supercontinent called Pangaea. Eastern North America was connected to Europe and Africa. During the late Triassic, ~200 million years ago, Pangaea started to break up and the continents began to move toward their present positions. The break-up of Pangaea resulted in stretching and finally fracturing and faulting of the continents near their margins. This type of faulting took place along the entire eastern margin of North America and resulted in the formation of large basins, one of which is the Fundy Basin. Basaltic magma intruded the crustal fractures forming dikes and sills and basaltic lava flowed out on the basin floors. Later, lakes formed on top of the basalt flows, and sediments were deposited on the lake bottoms, burying the basalts.

In Nova Scotia, these ancient lava flows now "cap" North Mountain between Cape Blomidon and Brier Island, a distance of more than 180 km, and are well exposed in bluffs along the Bay of Fundy shoreline. Post-Triassic crustal movement has elevated these rocks so that today they lie in slightly tilted layers with highest elevation ~240 m above sea level. The rocks tilt to the northwest, so that they extend under the Bay of Fundy. Where wave erosion has cut into the rocks, high bluffs have been sculptured out, and it is possible to study many features of the structure and mineralogy of relatively fresh volcanic rocks and overlying sedimentary rocks.

At Ross Creek, a section is exposed through the topmost lava flow of the North Mountain volcanic sequence, and into the overlying sedimentary unit (called the Scots Bay Formation).

STOP 1: The lower part of the flow

About 10 m from Ross Creek, you can stand on a flat, wave-eroded "platform" made of basalt. The basalt is composed mostly of two minerals, plagioclase and clinopyroxene. Because the lava cooled quite quickly, the crystals are small (1–2 mm in size), but on weathered surfaces you can see the elongate white grains of plagioclase, surrounded by darker coloured pyroxene.

At this location you are standing on rocks which formed in the lower part of a lava flow. As the lava cooled, shrinking occurred, to form structures called columnar joints. They are similar in form and origin to the mudcracks that you may have seen in mud puddles as they dry up after a rain. However, cooling lava is much thicker than drying mud, so the vertical cracks form column-like shapes—hence the name "columnar joints." These structures are not well preserved in this outcrop, but you can find some good examples. Viewed from the top, they form a polygonal pattern, and from the side they look like columns. As they weather and erode, they give the outcrops a knobby and hummocky appearance.

As you walk along the shore from Stop 1, you will notice "lines" (actually sheets) of lighter coloured material in the rock. These features are called veins; they are filled with minerals deposited from hot water solutions moving through cracks in the rocks. The minerals are mainly zeolites and chalcedony (fine-grained quartz). The zeolites are mainly pink and the chalcedony is grey, white, or red.

STOP 2: Minerals in Veins

About 40 m from stop 1, again on the wave-cut platform, you can see several (at least three) generations of cross-cutting veins, filled with different minerals (recognized by their different colours). You can trace some of these veins into the cliff. If you look at the veins in the cliff, you can see that the crystals have grown from the walls toward the centre of the veins, like teeth that do not quite meet in the middle.

Continue walking along the rocks on the upper part of the shore. Only one lava flow is exposed in this section, and it is the uppermost flow of the North Mountain sequence of 20 or so flows. As you walk from Ross Creek to the southwest along the shore you are seeing a section through this flow from near the bottom at Stop 1 to the top at Stop 3. Because of this, you will notice an obvious change in texture between Stop 1 and Stop 3.

STOP 3: The upper zone of the flow

As lava cools, gases trying to escape to the surface become trapped by a hard crust on the top of the flow, forming bubbles that accumulate in large numbers in the upper part of the flow. The gas slowly escapes, leaving holes called vesicles. Later, hot water moving through the rock may deposit secondary minerals in these holes, which are then called amygdales.

A variety of secondary minerals fill these amygdales, including many zeolites (heulandite, stilbite, and natrolite), chalcedony, opaline silica, and analcime. You can see many beautiful crystals, but they are very difficult to sample from outcrop or large boulders! For collecting, you are best to look for small loose pieces lying on the beach.

The lava near the top of the lava flow cools more quickly than that near the bottom—if you look carefully at the basalt in between the amygdales, you will notice that the plagioclase and pyroxene crysals are so small that you cannot see them with the naked eye, in contrast to those in the more slowly cooled basalt at Stop 1.

As you keep walking along the shore, you walk "through" the top of the basalt flow and into the rocks that were deposited about 190 million years ago as limy silt and mud on top of the basalt after it cooled.

STOPS 4 & 5: The Scots Bay Formation

The sedimentary rocks are light brown to pale yellow in colour, and are composed of limey siltstone. They are called the Scots Bay Formation, and reach thicknesses of several kilometers out under the Bay of Fundy.

The sedimentary rocks are mainly layered, but they contain massive lumps of another kind of rock called chert. These lumps are chert nodules. They formed when cavities were dissolved in the limey mudstone and hot water deposited fine-grained quartz (chalcedony) in the cavities. The chert varies in colour—the grey variety is commonly called flint, whereas the red variety is known as jasper. The chert is very hard compared to the limey mudstone, and forms "lumps" sticking out the the bank, and on the beach.

The boundary between the basalt and the sedimentary rocks of the Scots Bay Formation is called a contact. You can see the contact in the bank; it is tilted toward you. You can trace the contact under the beach and across the small cove to the bluff on the west side of the cove.

Walk across the cove by following the contact. You can recognize the sedimentary rocks by their layering and light colour, and the basalt by its dark colour and rubbly, amygdaloidal texture. Some large chert nodules are present in the sedimentary rocks exposed here.

About halfway across the cove, the contact changes tilt direction and emerges from under the beach to the bluff in front of you to the southwest. The basalt and overlying sedimentary rocks have been folded and form a downfold, or syncline, under the cove (see Fig. 3).

Above the Scots Bay Formation in the cove, the cliff is composed of poorly consolidated sediment. This material was deposited by glaciers as they melted away about 10,000 years ago. The common name for till is "boulder clay," and you can see why—it consists of material of many different sizes—many of the large cobbles and boulders are made of rhyolite and granite from the Cobequid Highlands on the other side of the bay—these rocks were carried over by glaciers. You can see that the till is easily eroded, and the cottage owners above must be very concerned as they watch their valuable land (and eventually their cottages!) wash away. The cottage owners on the bluffs made of basalt are in a better position—the rate of erosion there is much slower!

If the tide is low enough, turn and walk out toward the water. You are now walking back down into the basalt flow, and you will see that it is amygdaloidal. At very low tide, you can walk into the less amygdaloidal, lower part of the flow that you saw at Stop 1. Turn and look back at the cove. You should be able to recognize the features shown in Fig. 3.

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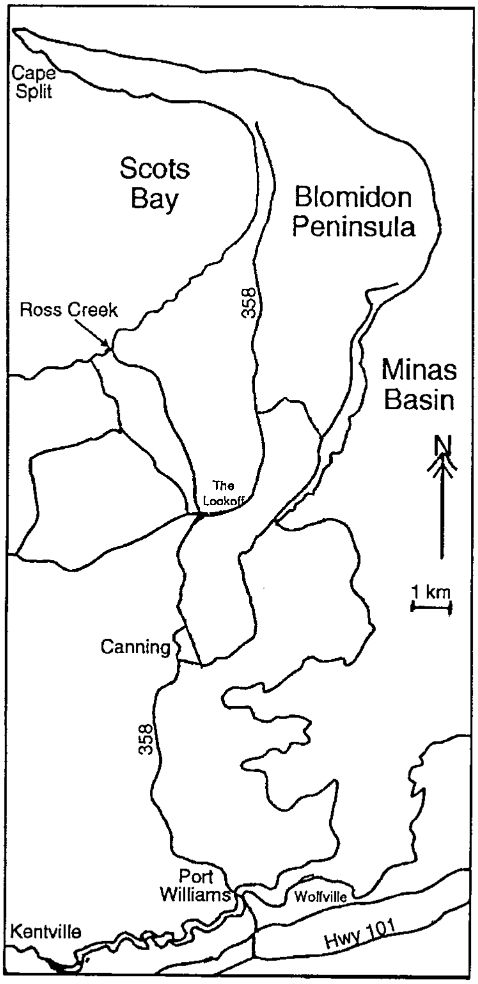

Figure 1: General road map showing location of Ross Creek

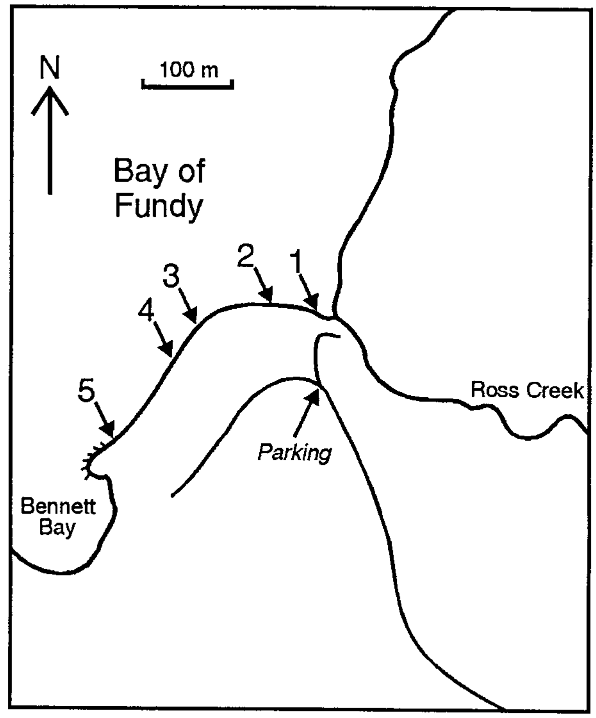

Figure 2: Location map of stops explained in the text.

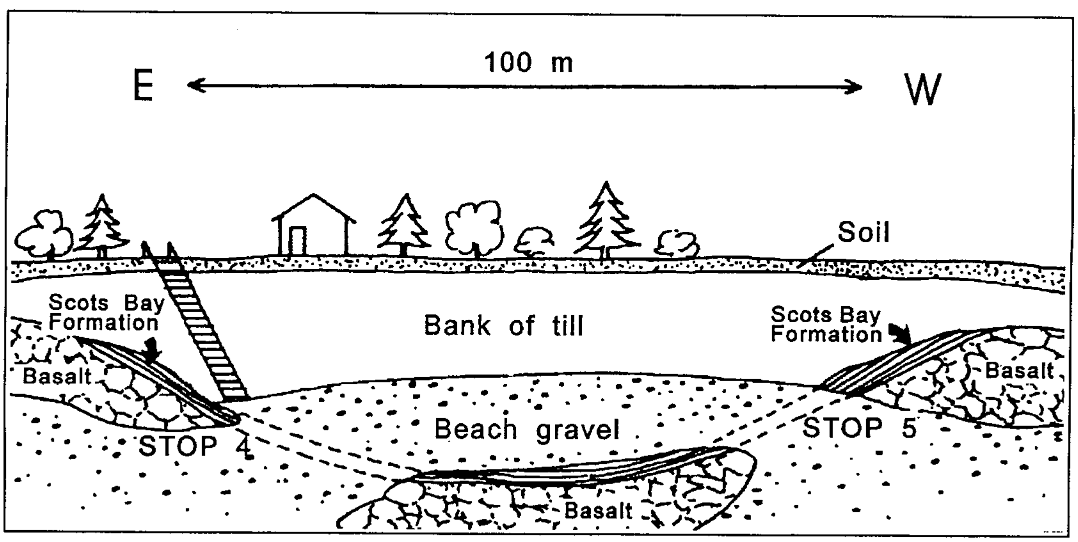

Figure 3: Detailed sketch of the geology in the cliff at stops 4 and 5.

Head of Saint Marys Bay

Quaternary Glacial and Marine Geology

Highlights

- submerged forest
- hummocky terrain
- wind gaps
- kettle hole in a glaciomarine delta

Location

This trip describes three sites along the head of Saint Mary's Bay off Highway 101 and out onto Digby Neck (Fig. 1). For the first location take Highway 101 southwest past the turnoff to Digby towards Brighton. At 5.5 km past the Digby turnoff, opposite a motel, is a dirt road going north towards North Mountain across a dyked salt marsh. If you are travelling northeast on Highway 101 it is 1.5 km northeast from the old wharf at the foot of the road to Bloomfield in Brighton. Take this road about 1.2 km to the houses at the end. Go down to the shore.

The second stop is in Centreville on Route 217 on the Digby Neck. To get there go back to Highway 101 and turn northeast towards Digby. At the next intersection, about 2.5 km, turn north towards Roxville. At Route 217 (approximately 4 km) turn west towards Brier Island. Go approximately 15 km along 217 to Centreville. In Centreville take the road to Trout Cove. While driving to the beach at Trout Cove look at the sides of the hills.

The third stop is 8 km further southwest along 217 at Sandy Cove. The small road to Burns Point follows along the outwash plain. Take this road to Burns Point and look at the boulders on the beach.

Introduction

This region is interesting because it exhibits an interesting sequence of Quaternary glacial and marine deposits overlying the bedrock. Quaternary glaciers effected the pre-Tertiary fluvial landforms of Nova Scotia through ice sheet and ice cap development and its latest effect has been variations on the relative sea level which in turn alters the shape of the embayments and causes important erosional and depositional cycles. At present the sea level is rising about 30 cm per century and the tides in the Bay of Fundy are growing.

This report describes late Wisconsinian glaciofluvial and glaciomarine and post-glacial marine deposits in the Saint Marys Bay region. They are typical of most similar deposits around the Maritimes and the coarser grained materials are locally important as aggregate resources and aquifers but have limited agricultural use other than as blueberry fields. In the case of the marine deposits they are important coastal wetlands and are great places to bird watch.

Descriptions

Head of Saint Marys Bay—submerged forest

Forested islands of till project from the plain, showing how the mud has buried the glacial landscape. An ancient forest can be seen on the old soil where shoreline recession has truncated a till knoll. In the surf zone within 3 m of high tide level, where wave action has prevented accumulation of modern shelly mud, several hectares of ancient forest are exposed complete with podsol and cover of brackish water (Scirpus) sedge peat marking the onset of marine inundation. A tree stump 2.6 m below high tide dated 2090 ± 70 y. B.P. (GSC-2911) gives the time of death by paludification whereas the first layer of brackish water peat marking the onset of marine conditions at -2.4 m dates 720 ± 130 y. B.P. (GSC-997).

Note the zonation of grasses at different tidal levels (*Spartina alterniflora* to mean high tide, *Spartina patens* to mean monthly high tide, and *Scirpus* spp and *Juncus gerardii* to mean annual high tide) and the nature of their remains. A section through 5 m of core to the basal forest soil shows many alternations of grass layers from different levels showing a fluctuating rise of sea level. A freshwater layer at -3.1 m indicates a pronounced regression.

Centreville—hummocky moraine

On the drive to Centreville note many the many hummocky terraines on which blueberry abounds. At Rossway the highway follows a large moraine trailing westward from a rock knob, apparently built by ice moving down Saint Marys Bay. Other moraines and cross-axial meltwater channels at Waterford support this hypothesis. Between Rossway and Waterford, in the valleys and on the slopes of hills, is a complex mixture of colluvium, derived from ice-frosted and shattered boulders, talus, mixed glacial and post glacial materials, and slumps.

At Centreville the wind gap to the Bay of Fundy (dirt road to Trout Cove) shows hillside meltwater channels marking the margins of the ice which flowed through the gap to the Bay of Fundy. Marine terraces abutting against the hummocky moraine, the highest of which is at +27 m AHT on the northeast side of the wind gap, show a high glacio-isostatic sea was in contact with the retreating ice.

Sandy Cove—glaciomarine delta

At Sandy Cove there is a glaciomarine delta considered to be one of the best exposed in the province. Goldwait (1924, p. 100) describes it as thus:

> This broad gap in North Mountain, instead of being swept by tides is blocked by a great gravelly plateau, whose flat top stands about 125 feet (26 m AHT) above the sea. On the Fundy shore, in a great 125-foot cliff, behind the beach, the stratified structure of the plateau is plainly shown. Sand is arranged in strata that dip south 50 degrees west (230°) at an angle of 20 degrees. This stratified structure looks like that of a deep delta built in the gap as the ice melted away from it. Near the middle of the gap this great plateau is interrupted by a huge hollow of irregular outline and steep sides. A pond occupies the bottom of the hollow; oddly-shaped kames rise around its borders, and one esker-like ridge projects far into it, nearly bisecting the pond. Plainly, this is a great ice block pit, or kettle-hole, marking the site of a mass of ice which was enveloped by the deposit. The preservation of the steep sides of the kettle hole, and of the kames around it, seems to require that the mass of ice lingered until the deposit had been elevated to its present position.

Another remarkable aspect is the lithology which is not of the surrounding basalt, nor Triassic red beds from the floor of St. Marys Bay, nor greywacke from Southern upland, but quartzose material. This may derive from the interior granite, but its absence in an up-ice direction such as on the coast at Weymouth argues against such a provenance. Alternatively it may be the most resistant fraction from the pebbly Blomidon Sandstone under St Marys Bay, or finally it may have been recycled from some younger deposit such as marine gravel of Quaternary age.

The age of the feature is not known but it must predate the 14 000-year old marine beds at Gilbert Cove. Shells were reported but none are known now. It most likely dates from the final radial movement of Nova Scotian ice, and as it is near the outer limits of ice-flow indicators left by that movement, the delta may mark the maximum extent during the Late Wisconsinian Stade.

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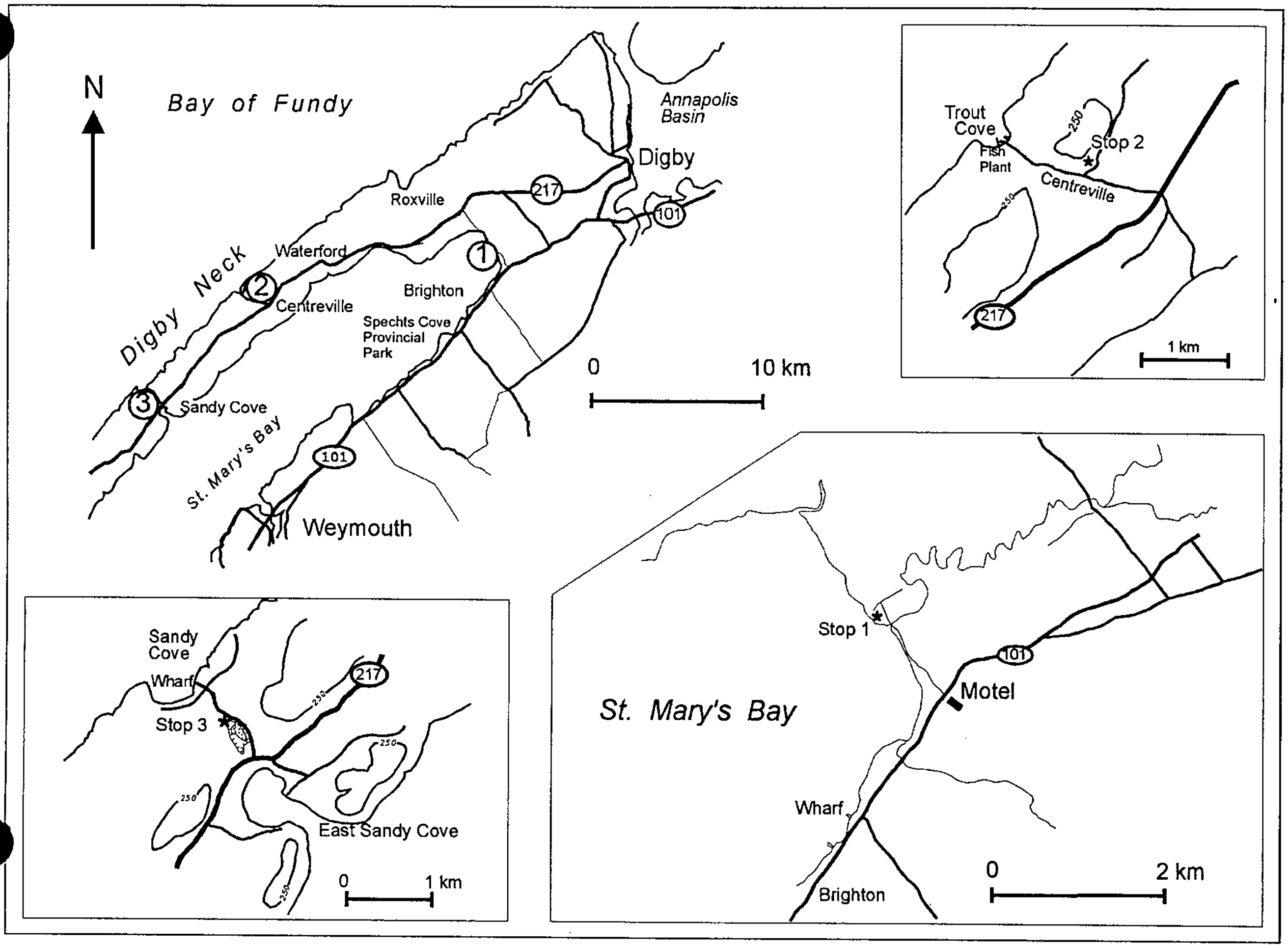

Figure 1: Location maps of stops 1–3. Line in figures for stops 2 and 3 are topographic contours.

Cape St. Marys

The Contact between the Halifax and White Rock Formations

Highlights

- contact between Halifax and White Rock Formations
- diamictite of the Halifax Formation
- microfolding in the Halifax Formation

Location

At the town of Cape St. Marys on the Bay of Fundy, Nova Scotia (Fig. 1). Take Highways 101 and 1 west to Meteghan and then continue southwest on Highway 1 to Mavillette. Alternately, if you are coming from Yarmouth, take Highways 101 and 1, north, and at Woodvale take the exit off Highway 1 to Salmon River and Mavillette. Take the road west from the north end of Mavillette to the village of Cape St. Marys. Drive out to the lighthouse or park near to the wharf and walk to the cape.

Precautions

The field trip involves a coastal section in the Bay of Fundy with one of the highest tides in the world. The tide does not pose a hazard at this location but in stormy weather, high tide may be higher than usual. Also the rocks are very slippery near the tide line, as well as when they are wet. Please exercise extreme caution and common sense when viewing this section.

Introduction

This is one of the best, although controversial exposures of the contact between the Halifax and White Rock Formations. The contact has been interpreted as conformable by Schenk (1972) with the top of the Halifax being a glacio-marine tillite but recent work by Culshaw and Liesa (in prep.) interprets the contact as an unconformity as initially proposed by Taylor (1969). On the Geological Map of Nova Scotia (Keppie 1977) the contact is depicted as a thrust.

The contact is exposed in two areas, 500 m along strike. The best exposure, at the cape (Fig. 2) and described here, is part of a continuous section hundreds of metres thick starting in the Halifax Formation and extending about 90 metres into the White Rock. The other section along the beach (located on Fig. 1) begins briefly in the Halifax and extends further into the White Rock, but is not described here.

The section is within the Cape St. Marys Shear Zone of Culshaw and Liesa (in prep.), one of many shear zones associated with post-main phase (i.e., post-Acadian) deformation. The main phase deformation produced southwest trending, upright and horizontal folds with associated axial planar cleavage resulting from the docking of Laurentia with North America. When Gondwana collided with Laurentia and the other terranes to the west, post-main polyphase deformations modified these earlier formed structures to what we see today at Cape St. Marys.

Description

At Cape St Marys, the exposures along the cliffs show a very thick, structurally disturbed section of the Halifax Formation overlain by some 110 m of the lower part of the White Rock Formation (Fig. 2). The Halifax consists of metamorphosed mudstone with very continuous, even, although very thin, silt layers containing parallel laminations. Cross lamination is rare. Much of the layering is transposed bedding (composite S_0 and S_1) coplanar to a second cleavage (S_2) leaving top indicators non-existent. As well, many of the silt layers show small open to isoclinal folds with post-main phase crenulation cleavage (S_3) that is ubiquitous throughout the section and is generally steep and axial planar.

The top of the Halifax is a diamictite or pebbly slate, 20 to 100 cm thick. It has abrupt contacts with the underlying slates and overlying felsites. This unit has been correlated to a similar lithology at Elderkin Brook, south of Kentville, 230 km to the east. There it is 35 m thick and occupies the same stratigraphic position between the two formations.

The White Rock Formation above this contains a very wide range of lithologies (Fig. 3)—felsite tuff, greenstone, quartz arenite and wacke, meta-siltstone, slate, greenstone meta-basite. Volcanic rocks are concentrated in the basal 20 m, quartz arenites over the middle 60 m, silty slates over the upper 25 m, and greenstone meta-basalt at the top of the section. The base of the section shows a transitional contact (1-2 cm thick) between the Halifax and White Rock—the colour of the slates blanch, and primary lamination is lost. Local, slight angular discordance is either ambiguous or superficial. In the diamictite and White Rock Formation, the bedding parallels the vertical cleavage.

Discussion

There are three problems presented by these rocks: the nature and origin of the meta-diamictite in the upper Halifax; the nature of the contact itself; and the nature of the deformation. Absolute and unquestionable evidence to the solution to these problems does not exist so the observer is free to support any or all interpretations. Following is a brief description of the problems and their implications.

Origin of the meta-diamictite

The diamictites at the top of the Halifax Formation possess the following characteristics:

(1) Unsorted, gravel-sized (1–5 mm) to cobble-sized (up to 30 cm in diameter) clasts are matrix-supported within arenaceous schists and argillites. The average size of clasts is 1–5 cm.

(2) The matrix is the metamorphosed product of a sandy mud with 30–50 per cent silt, along with intermixed sand and granules.

(3) The clasts are very angular, including some rod shapes. A few clasts are faceted and flat-iron shaped. Striations on the clasts do not correspond with cleavage and microfracture orientations (Schenk 1972).

(4) The lithology of the clasts is meta-sedimentary including meta-arenite, meta-siltstone, slate, felsite, and chert. Very fine-grained arenites and siltstones are the common rock type (75 per cent of all clasts). Except for the slate clasts, these lithologies are unknown in the interstratified or underlying slates.

(5) The rock has an unusual fabric. Clasts are concentrated in small lenses (5–20 cm high by several metres wide) and are dispersed singly along bedding planes. These lenses are intercalated with poorly sorted argillites or schists (gravelly sandy mudstone). Beneath single clasts, slate laminations are deformed and broken, similar to dropstone fabrics.

(6) At Cape Saint Marys, the diamictite unit is one metre thick and lies immediately below felsic volcanoclastics with intercalated lithic tuffs and ash flows.

These meta-diamictites possess the characteristics of pebbly mudstones—unsorted clasts supported by a poorly sorted very fine-grained matrix. Mass flow deposits and tillites have these characteristics (Dott 1961; Walker 1976). Since the clasts are very angular, exotic, and prelithified, they would have been eroded from a bedrock mass, by either mass movement or glaciation. They have the following essential features of glacio-marine deposits: a) faceted clast shapes that are typical products of glacial erosion; b) dropstones (Schenk 1972), diagnostic of glacio-marine sediments (Frakes and Crowell 1975); and c) dropstones and small lenses (clusters), typical marine drift ice deposits (Dangeard and Vanney 1975). They lack typical mass flow features such as grouped clasts, slumping, large boulders, and blocks. There is no associated nearby slope necessary for mass movement, i.e., they are found intercalated within slatey units.

The assemblage of structures and textures along with regional associations and correlation with other unequivocal glacially derived units has led Schenk (1972) to interpret this deposit to be of glacial origin. The source rock for the clasts is unknown but is considered to be distant and the clasts were presumably transported to the area by ice rafting.

The Contact

The regional configuration of the Goldenville Halifax-White Rock-Kentville sequence is one of conformable continuous sedimentation from Ordovician to Devonian time with possible regional local unconformities between the Halifax and White Rock. At Cape Saint Marys, the type of contact (conformable or angular unconformity) is complicated by later deformation making interpretation ambiguous. The layering in the Halifax is truncated at a low angle by the diamictite and White Rock lithologies (Fig. 3) and the post-main phase deformation in the Halifax is polyphase, coplanar to discordant. Deformation in the White Rock is polyphase and mainly co-planar.

Schenk (1972) interprets the contact as conformable explaining the angular discordance being the result of tectonic attenuation of the less competent Halifax against a more competent White Rock lithology. Keppie (1977) interprets late deformation as obscuring any evidence clarifying the type of boundary so the author shows the contact as a thrust (Keppie 1979). Recent shear model studies by Culshaw Liesa (in prep.) imply initial conditions of an angular unconformity similar to that proposed first by Taylor (1969). They suggest the boundary was initially folded during mainphase deformation and then later reactivated as a thrust plane.

The post depositional deformation

All the foregoing descriptions of lithologies and structures are complicated by the deformational history. Fig. 3 shows detail of the structures seen at the Cape Saint Marys. The layering is transposed bedding (composite S_0 and S_1) and co-planar to S_2. Several antiforms with associated micro-folds (F_3) and penetrative crenulation cleavage (S_3) occur in the Halifax Formation within 50 m of the contact. The structural style changes relative to distance from the contact. Within 5 m of the contact the limbs of these microfolds are attenuated, broken, and refolded. Many of them may be sheath folds indicative of high strain. The clasts of the diamictite have been stretched out into a general ratio of 6:1:2 (Schenk 1972) and these along with pressure shadows indicate southeast-side-up (reverse sense) shear. At the contact and into the White Rock, attitudes of stratification (S_0) revert to the vertical position, parallel to the cleavage (S_2 in the White Rock and S_3 in the Halifax). The steeply dipping strata of the White Rock Formation are penetratively deformed as well, and the prominent cleavage (S_2) has the same vertical attitude as the bedding.

This whole package has been broken up into small blocks by local faults, generally east-southeast trending and dipping steeply to the east. Because of the lack of marker beds, throws on the faults are difficult to determine. Jointing is evident with major trends being north-south and northwest-southeast.

Metamorphic grade is low in the chlorite zone. In the Halifax Formation, away from the contact and intense deformation, pressure-solution cleavage is evident. Overall, clastic textures are preserved in both the Halifax and White Rock Formations.

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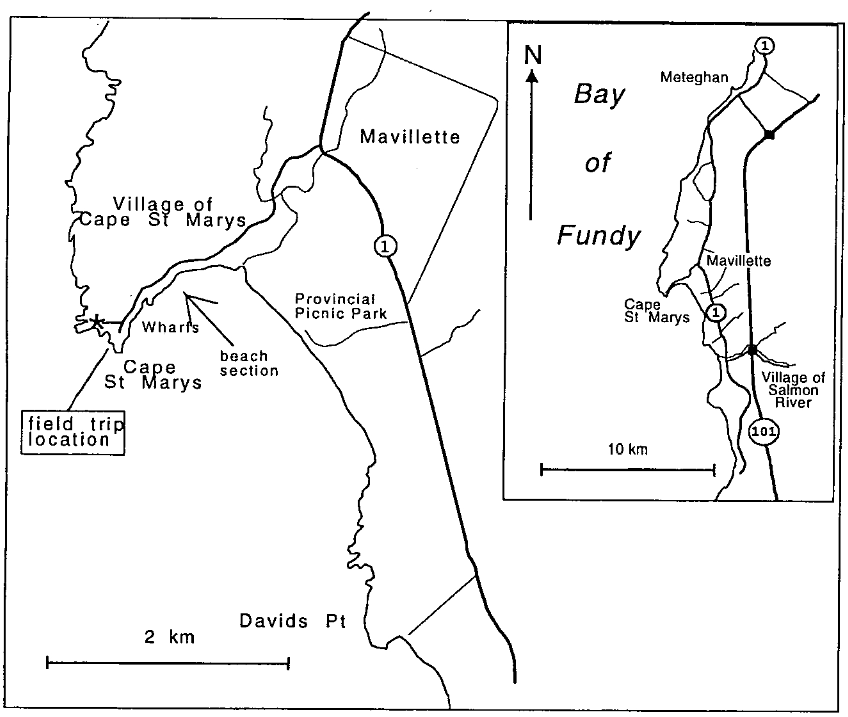


Figure 1: Location map.

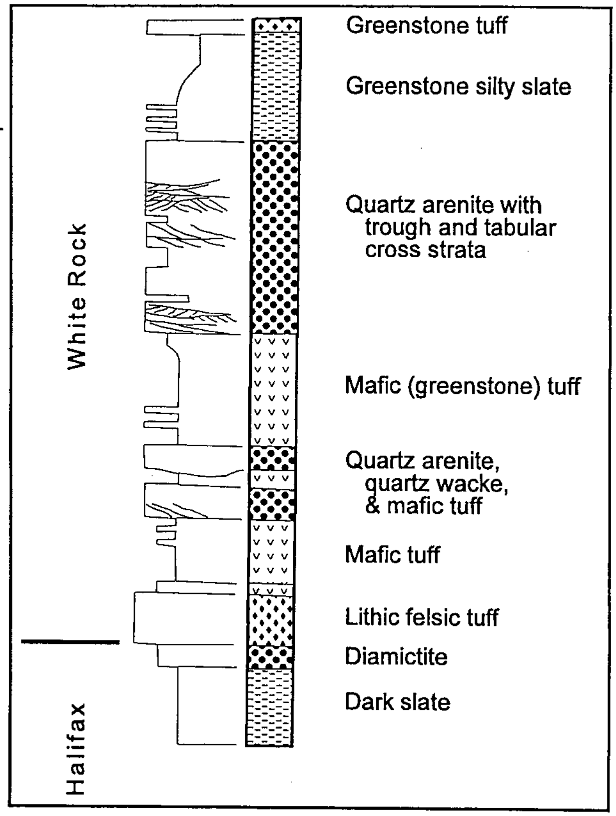
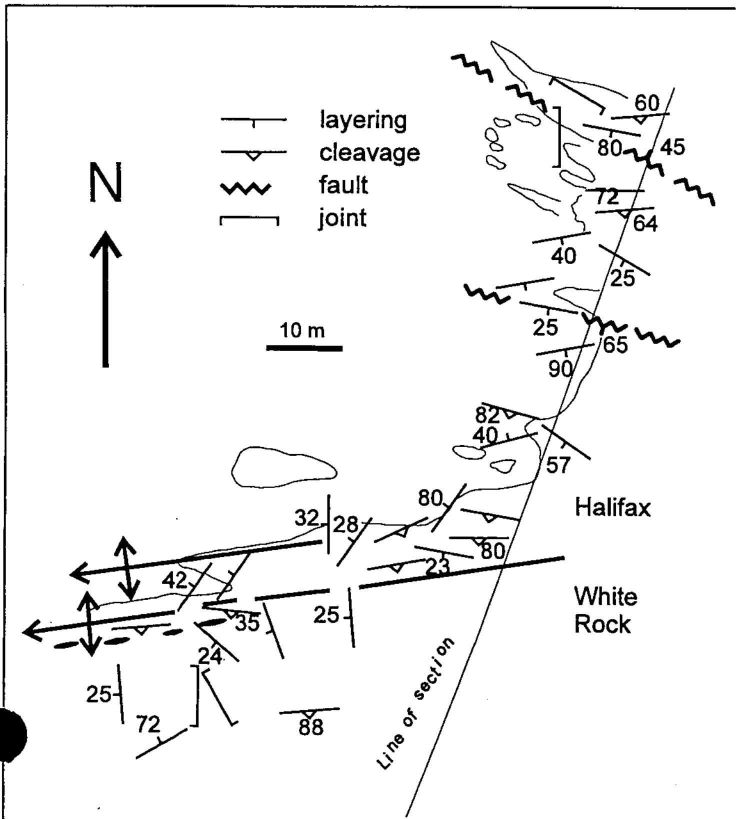


Figure 2: Generalized stratigraphic column for the Halifax–White Rock transition at Cape Saint Mary. No scale. From Schenk et al., 1980.



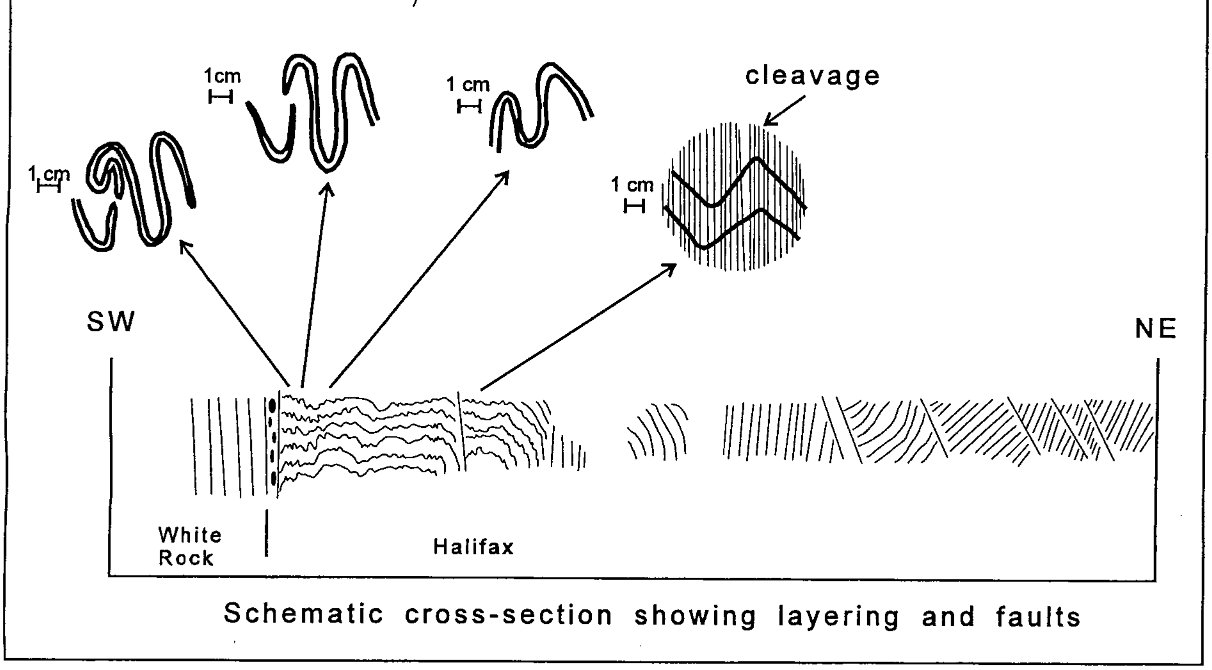


Figure 3: Detailed geology at Cape Saint Mary.
From Schenk et al., 1980.

Salmon River

Multiple Till Deposits

Purpose

The shore along the Bay of Fundy just north of the Salmon River has some of the best exposures in Nova Scotia of multiple till sheets.

Precautions

These till sections are unconsolidated deposits along the shore of the Bay of Fundy and as such are subject to sudden weathering and erosion. What may be exposed one year can be covered, eroded, and lost the next. Also, the sections are unstable so climbing is not recommended. The intertidal and parts of the beach area are extremely muddy and rubber boots are recommended. Be aware of the state of the tides at all times.

Location

Take Highway 1 north and go across the Salmon River (Fig. 1). One kilometre beyond the river is a church with a side road going to the west, towards the shore. This is David Point, the site of the first till description and a type section for the Pleistocene glacial deposits. The second description is of the Cape Cove regional reference section. It is along the shore from the Cape View Motel and Restaurant and extends south from the Mavillette Beach Provincial Campground boundary.

Introduction

The glacial history of Atlantic Canada cannot be compared exactly with the classical mid-continent history due to the lack of correlation of beds between the two areas, lack of ages, and lack of chronology through pollen spores and ages of marker horizons. Therefore Atlantic glacial geology has developed a local succession largely produced by a separate glacier domain composed of several independent ice centres of maritime type which were satellite to, and only locally in contact with the Laurentide Ice Sheet. In southwest Nova Scotia, the Yarmouth to Digby successions give a fairly complete succession of local glacial episodes and it is from this that the following scheme is proposed.

The oldest deposits may be the intensely oxidized deposits of the Bridgewater and Mabou Conglomerates and they may predate the earliest glaciation. Pre-Illinoisan interglacier deposits are indicated from dated shells in till in southwest Nova Scotia and the last interglacial, the Sangamonian (75 000-128 000 y. B.P.), is represented by marine and terrestrial deposits underlying Wisconsinian tills.

It is not known when glaciers first developed in the Wisconsinian but the first major flows were first eastward followed by southeastward. In Cape Breton this started about 50 000 y. B.P. but were probably earlier in the west. The second ice phase was southward from the Escuminac Ice Center in the Prince Edward Island region. This was followed by a third phase during which the ice center shifted to an ice divide in southern Nova Scotia resulting in flow directions to the north (in the north and west along the Bay of Fundy from Yarmouth to all of northern Nova Scotia) and south (in the Halifax region) across this divide. The final phase resulted from the breakup of this divide into remnant ice caps that were centered on highland areas. During this phase the ice moved generally westward into marine areas.

Description: David Point

The type-section for Pleistocene glacial deposits is exposed along the shore at David Point, near Salmon River. Here there are five major units (Fig. 2) reflecting a sequence of old till with interglacial material, weathered middle till(s), and immature surface till.

Unit 1—A red-brown till (Red Head Till) with many shell fragments and foreign stones that, in places, sits directly on top of a grey metasedimentary rock. Where truncated and exposed in the intertidal zone, the till exhibits a strong south (185°-190°) fabric, with platy clasts on edge.

Unit 2—A partly oxidized fossiliferous sand (Salmon River Sand), in which contact with the enclosing tills is conformable and transitional as evidenced by small-scale interbedding. Projections and lenses of sand are seen in the lower till and a lens of brick-red stony pelite, in every respect like the lower till, is found in the upper third of the sand bed, emplaced as if by flowage. These intimate relations indicate ice recession in a deep transgressing sea, where conditions were suitable for molluscan life, followed by renewed subglacial till deposition. If the brick-red lens is a flow till, the ice front during sand deposition was near, perhaps against the back slope of the till mass itself. Sand laminae in the red pelite are composed of coarse well-rounded frosted grains suggesting a nearby beach environment.

The molluscan assemblage is remarkable for its variety, excellent preservation and warm-water affiliation. One gastropod stands out as extinct and yields a Carbon-14 age of 38 600 ± 1300 y. B.P. (GSC-1440). The finite mid-glacial age is corroborated by Uranium-Thorium (33 000-40 000 y. B.P.; Lamont-Doherty-1348a) and amino acid racemization (70 000-80 000 y.B.P.) dating. The paleo-environment of this fossil and the associated fauna was most likely interglacial in the sense that water conditions were comparable to those in the area today. They have been correlated with beds across the Gulf of Maine in New England that are thought to be of cool-temperate Wisconsinian age. There is thus the difficulty of explaining glacier-marginal marine deposition in a sea of interglacial character. Significantly, in the western Gulf of Maine, during the late glacier prior to 14 000 years ago, temperate water foraminifera lived below an ice shelf. Perhaps during glacial-age low sea levels, an emergent Grand Bank deflected the cold Labrador Current far out to sea, thus permitting the warm Gulf Stream to penetrate the Gulf of Maine.

Unit 3—A 6-10 m thick light grey, sandy till of which the upper 4 m (Saulnierville Till) is distinctly browner due to iron accumulation. In it are sub-horizontal sorted horizons and stone lines, but shell fragments are rare.

Unit 4—Surmounting the south end of the section is a lens of olive-grey sandy rubbly till (Beaver River Till), like the surface till seen further to the south of the area. It is thought to be an end moraine marking the limit of the ice advance over the older tills.

Unit 5—A marine gravel which was deposited after the glaciers retreated. It rises from about 8 m above tide level on David Point, where it is 6 m thick, to a point just south of the road where it pinches out at an elevation 11±0.5 m (marine limit) against a rolling surface of till. Internally, sandy gravel foreset beds grade to silty bottomset beds and the whole is truncated by planar topsets. The wedge grades up to the hummocky moraine of young till and is therefore interpreted as proglacial marine outwash. The gravel is not developed on the inland or proximal side of the moraine, proving that ice and sea were in contact while there was glacio-isostatic regression.

Bedding structures are a crude measure of former tidal range. Amplitude of the foresets here indicates a maximum range of 2–3 m, whereas present range exceeds 9 m. Tidal amplification is part of the reason for the rapid recent submergence measured in estuarine accumulations.

Description: Cape Cove/Cape Saint Mary

Along the shore at Cape Cove are exposed ten units that span the last glacial cycle (Fig. 3). The section is about 1.5 km long and gently dips to the north so that a walk along the beach to the south is a trip back through the last 100 000 years. The units are described below in order of increasing age to reflect this walk. The distances are metres south of the park boundary and are approximate only.

Unit 10—Holocene sediments

The road to the beach overlooks a broad tidal marsh built up during the Holocene rise of sea level. The latter marine phase is contrasted with earlier postglacial submergence and regression that produced a mantle of sandy gravel by reworking of underlying units. Its sorting, roundness, imbrication, and planar bedding are characteristic of the littoral facies. As elsewhere the lag deposit is not more than 2 m thick, and is composed of loose unweathered material. It pinches out upslope at distance 430 m at a height of 13.5 m (marine limit) that is manifest on the surface as a low, but distinct notch.

Units 7 & 8—Young (= Late Wisconsinian) Surface Till (Beaver River Till)

Till corresponding to the glaciation immediately prior to the glacio-isostatic transgression is not found either below or in lateral contact with the gravel, but occurs farther south at 1100 m to 1300 m where it is the topmost in a sequence of four tills as a 2 m loose rubbly slate mantle with a half-metre of soil underlying an area of hummocks with associated proglacial meltwater channels. It is thus possible that the late Wisconsinian ice margin lay inland of the present coast. Detailed mapping is needed to substantiate this.

Unit 9—Cemented Fluvial Gravel

The young, reworked surface gravel contrasts with its parent material, a conglomerate of sandy gravel cemented by dark brown iron hydroxide which outcrops at beach level between 80 m and 50 m. Some involutions or cryoturbations are seen at 145 m, but the remarkable feature is that some beds are tilted to vertical to overturned, and two isoclinal folds and a thrust fault, all directed to the south, are present. Glaciotectonic, mainly plastic, deformation by south flowing ice (prior to cementation) was the probable cause. Oddly some beds are composed of fresh grey uncemented gravel at 250 m and 450 m. All structures indicate a subaerial fluvial environment: cross-bedding, scour channels, and imbrication. The stratification dips

24° northward, and the cross-stratification is inclined 32°-35°. Both are too steep for normal fluvial bedding (the angle of repose of loose, dry material is 35°) so some general tilting of the whole mass is suggested and indeed, evidence of up-to-the-south faulting is seen at 800 m. The deposit is thought to be marine outwash from a nearby ice-front, graded to à lower sea level. Till from the glacier responsible for the gravel, and occupying the same stratigraphic interval, is seen at 800 m to 1200 m but the contact relations are obscure and the correlation remains hypothetical. The gravel rests conformably on a smooth surface, apparently truncated across a conspicuous bed of Unit 7.

Unit 6—Reddish-brown Shelly Till

This massive body, about 15 m thick, emerges from the beach level at 500 m and rises southward over a complex of older tills and gravels. It has the same properties as elsewhere in the region: anomalous red colour, New Brunswick erratics, compaction, southward-oriented fabric (210° alignment of vertical clasts where truncated in intertidal zone; 180° stossing on large boulders in basal part; 170° striations on underlying bedrock), and abundance of mollusc shells. These are mainly broken, but stout species like *Mercenaria* and *Verericardia* can be found intact. The intense colour obscures possible darkening due to weathering, and there is no sign of iron stain where it is overlain by grey till at 900 m.

Unit 5—Lower Light-grey Slate Till

Between distances 700 m and 900 m, where bedrock forms the lower part of the cliff, the stratigraphy becomes more complex: two tills lie above the reddish till, and one, together with three other non-glacial members, lies below. The till is 2–4 m thick, and crops out, partly over bedrock, along a distance of 550 m. It is greyish and mainly composed of local rocks and clearly indicates the activity of local ice prior to the arrival of regional ice.

Unit 4—Colluvial Debris

Between 725 m and 1000 m the lower till overlies a bed up to 4 m thick, composed of angular cleavage fragments of the local slate, that pinches out up-slope against bedrock. The clasts are imbricate downslope like talus, so the bed is interpreted as an apron of frost-riven debris that was probably emplaced during the periglacial conditions which preceded the arrival of glacier ice in the last cycle. Underlying sediments support this view.

Unit 3—Rusty, Oxidized Gravel

The slate colluvium conformably overlies horizontally-stratified yellowish gravel composed of well-rounded, conspicuously weathered cobbles that reach an elevation of 4 m above present tide level. Discoid clasts are imbricate, indicating onshore water movement. It is evidently marine in origin and probably dates from the last interglacial period.

Unit 2—Emerged Platform and Conglomerate

From 700 m to 1000 m a fossil wave-cut platform is being exhumed from the cover of sediments described above and is being stripped of its characteristic hematitic film. It rises southward from 1 m below tide level to just above tide level at 980 m where it is backed by a fossil sea cliff. It is odd that it ranges only one to two metres above its modern counterpart. At several points the weathered beach gravel rests upon it. At 900 m a small thin carapace of hematitic slate-shingle sandy conglomerate (beach rock) is cemented to the platform. Mainly local rocks comprise the deposit, in contrast with the more varied lithology of the overlying leached gravel. Nonetheless, the conglomerate is tentatively regarded as the basal part of the old beach gravel where translocated iron has accumulated.

Unit 1—Postglacial Faulting of Bedrock

The glacially-bevelled outcrop of slate at 800 m shows regular up-to-the-south displacement along its vertical cleavage planes. The heave averages 5 cm on more than 20 dislocations across a width of 3 m. The net movement is more than one metre and is also vaguely apparent as an offset of the overlying grey shaly bed. The outcrop is overlain by lower grey till and individual striations are traceable across the steps, proving that the faulting occurred after the earliest glaciation.

Postglacial displacement of vertically cleaved rocks in the region has long been known and it is a common phenomena of glaciated Appalachia. The movement is therefore tentatively interpreted as normal postglacial rebound that, rather than being a smooth warping, is concentrated along incompetent zones such as slate belts.

Source

Grant, D.R., 1980. Quaternary Stratigraphy of Southwestern Nova Scotia: Glacial Events and Sea-Level Changes. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual meeting, Halifax '80; Trip 9, Guidebook, 63p.

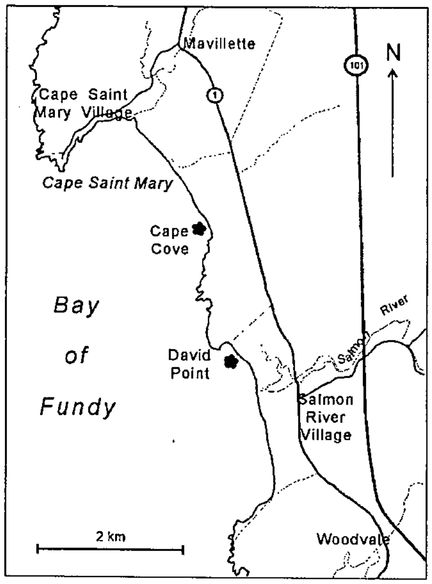


Figure 1: Location map for Cape Cove and David Point

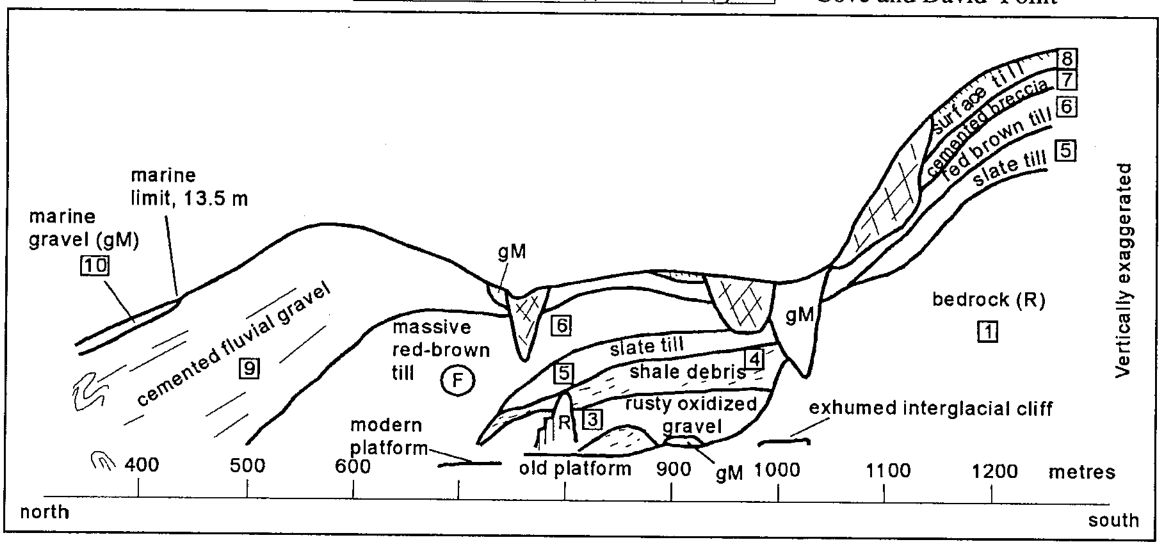


Figure 2: Multiple till section at Cape Cove. From Grant, 1980.

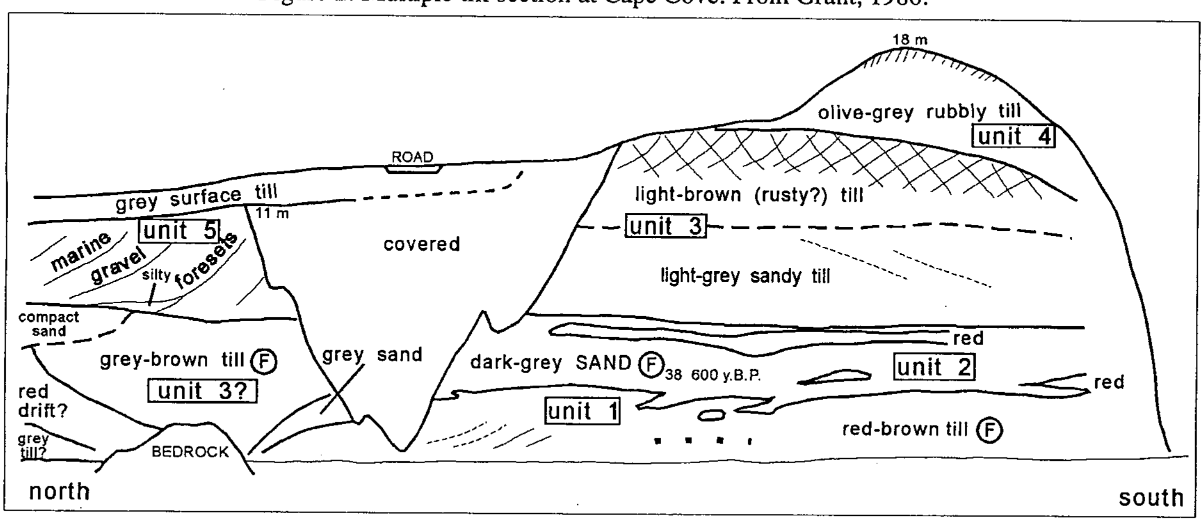


Figure 3: Multiple till section at David Point. From Grant, 1980.

Shelburne Area

Staurolite-Andalusite Schist at Jordan Falls and Rocks of the Shelburne Pluton Margin

Highlights

- staurolite-andalusite schist
- granodioritic to granitic plutonic rocks
- garnetiferous pegmatite and aplite dykes

Precautions

Both stops are along the busy Highway 103 (Fig. 1) although you can get off onto a dirt track at Jordan Falls if you so desire. Jordan Falls is a series of low glaciated outcrops but the Shelburne area has unstable, medium-high road cuts. Do not step onto the roadbed; this means stay next to the outcrops. The cliffs are dangerous and best viewed from below. The Shelburne stop is tricky because other drivers may think there has been an accident and will probably stop as well; maybe you should post a large lettered sign in the back of your vehicle saying you are on a field trip and looking at the rocks.

Introduction

Most of southern Nova Scotia is underlain by the Meguma terrane composed of the Meguma Group, a thick package of variably metamorphosed Cambrian-Ordovician sandstones and shales, and younger (Silurian to early Devonian) slates, quartzites and volcanics intruded by mainly peraluminous, granitic plutons. The dominant structures within the Meguma terrane are tight to isoclinal folds, with wavelengths of 10 to 20 km and axial traces trending northeast-southwest to more southerly in southern Nova Scotia. Metamorphism ranges from lower greenschist (or sub-greenschist) facies to upper amphibolite facies with the pressure of metamorphism being uniformly low. The highest grade rocks are exposed in the southwestern part of the terrane, in the Shelburne area, and at the eastern end in the vicinity of Canso. In between these two areas the metamorphic grades are relatively low.

The metamorphism which culminated in the development of staurolite and andalusite-bearing assemblages in the Shelburne area was part of a regional event that affected the entire Meguma terrane. $^{40}Ar/^{39}Ar$ dating has yielded ages in the range of 405-365 Ma (405-390 chlorite, 365-380 biotite and muscovite, and 362-387 micas in hornfelsic rocks; Muecke et al. 1988 and Reynolds et al. 1981) but dating the high-grade staurolite-andalusite metamorphism remains elusive due to overprinting by an Alleghenian event involving shearing, local mineralization and hydrothermal alteration. Plutons (350-320 Ma) appear to cut the high-grade regional metamorphic rocks so their age of formation was most likely early around 400 Ma. A late (400-375 Ma) and weaker metamorphic event which resulted in the scattered development of cordierite also preceded the plutonism.

This trip involves two roadside stops. The first is to view the high grade metamorphic rocks at Jordan Falls. The second is to look at the margin of one of the plutons in the Shelbourne area with its associated pegmatitic, aplitic, and granitic dykes.

Jordan Falls

Go west 1.7 km on Highway 103 starting at the bridge over the Jordan River. On the south side of the road is a small dirt track with a series of glacially scrapped outcrops in a clearing.

At this point the outcrops are medium-grade metasedimentary rock, generally of low to moderate dip on a NNE strike. They occur in the trough of a syncline, near the base of a thick pelitic sequence. Precise correlation with units in the Goldenville-Halifax transition is not possible because of the distance and the known lateral variations in the transitional members. However, the abundant examples of (now deformed) ripple cross-laminations in sandy layers and the range of compositions among the layers suggest affinity to the Risser Beach member of the Goldenville Formation.

The metasediments contain poïkiloblasts (small grains of one mineral within larger recrystallised grains of another mineral) of staurolite, garnet, andalusite, cordierite, biotite, and magnetite. Biotite and garnet grains tend to be small and are incorporated into the other metacryst minerals. Staurolite is typically about 1 cm across, and weathers out to produce idioblastic-staurolite gravel (large minerals bound by crystal faces), with many examples of interpenetration twins. In thin section staurolite porphyroblasts are inclusion-free. Randomly oriented andalusite metacrysts reach up to 15 cm in length. They occur as vague, pale pink patches on smooth surfaces, and are poikilocrystic, with up to 95% inclusions in the sandier layers. Elsewhere, andalusite has replaced matrix plagioclase and muscovite, but quartz, garnet, and opaque minerals have remained inert, and, as inclusions, show the same distribution in the andalusite as in the matrix. Cordierite metacrysts appear to have formed at least in part by replacement of andalusite, although there are many examples of hexagonal cordierite patches (now altered to pinite).

Because growth of these metacrystic minerals is governed by the availability of aluminum, iron, and magnesium, they are best developed in pelitic beds. Graded bedding is therefore reflected in the growth of new metamorphic minerals, but with the grading reversed. Some layers are particularly rich in certain minerals, reflecting subtle variations in the original compositional layering. Some particularly staurolite-rich layers, 20-30 cm thick, can be seen in the outcrops nearest the road.

Shelburne Pluton margin

Proceed 8.3 km west from the previous outcrop at Jordan Falls, 1.5 to 2.0 km west of Exit 26 from Highway 103, over the brow of the hill, on the north side of the road.

The eastern contact of the Shelburne Pluton is indistinct. Across a distance of about 2 km, the abundance of granitic and pegmatitic dikes gradually increases. The boundary of the pluton is arbitrarily placed where the ratio of plutonic rock to meta-sedimentary rock is 1:1. West from that point for a further distance of 4 km, screens and xenoliths of country rock become decreasingly abundant in the pluton. By this definition, this outcrop is part of the Shelburne Pluton, although metasedimentary rocks are still abundant, especially in the eastern part.

The Shelburne Pluton consists of muscovite-biotite granodiorite gradational to granite and tonalite. It is typically medium-grained, equigranular, and foliated. It has been intruded by garnetiferous pegmatite and aplite dikes, and contains abundant meta-sedimentary xenoliths. Recent $^{40}Ar/^{39}Ar$ dating suggests a Devonian age (370-340 Ma)for the pluton, essentially coeval with the South Mountain Batholith (about 370 Ma). The $^{40}Ar/^{39}Ar$ age spectrum is disturbed, like those from other plutonic rocks in southern Nova Scotia, reflecting an Alleghanian event which affected this area at about 320-300 Ma.

Contributor

Raeside, P.R. and Jamieson, R.A. 1992. Low-pressure metamorphism of the Meguma Terrane, Nova Scotia; Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Wolfville '92; Field Excursion C-5, Guidebook, 25 p.

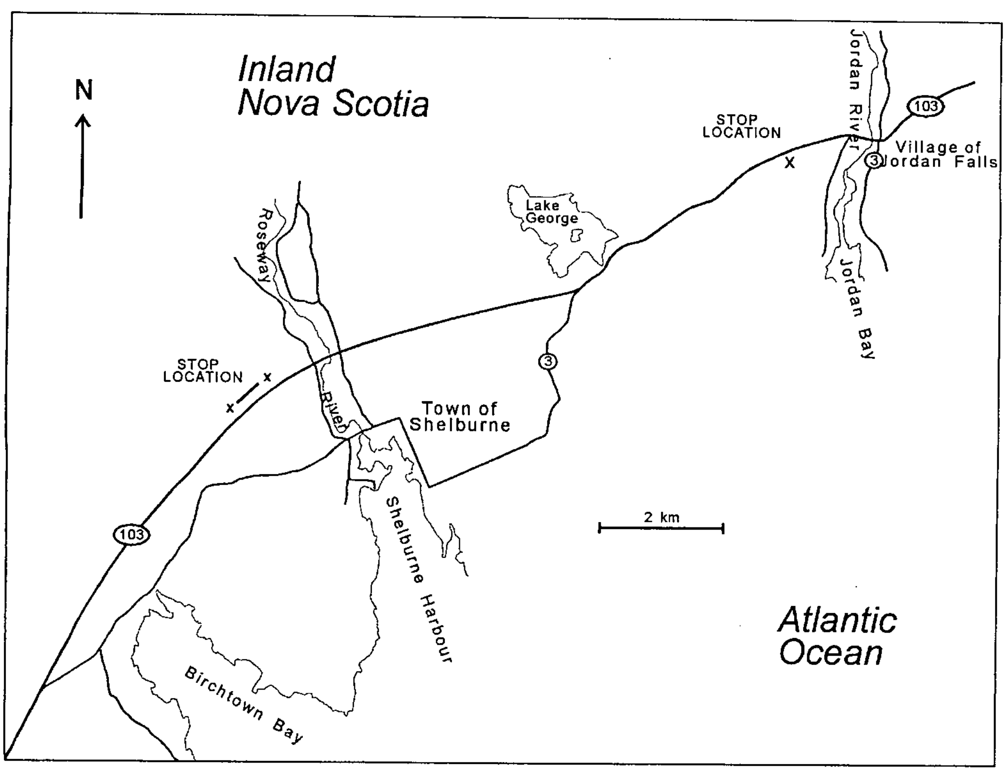


Figure 1: Location map for Jordan Falls and Shelburne field trip stops

The Ovens

Gold Bearing Veins and Their Relationship to the Structural Geology of the Meguma Group

Purpose

Coastal exposure providing a cross-section through the Ovens Anticline. Perhaps the best place to see fold-related structure and auferous veins in the Meguma group.

(Note: The area described occurs along coastal sections, access to which is limited by tides. It is strongly recommended that you correlate your visit with low tide; tide information provided in local newspaper.)

Location

The Ovens Natural Park is located on the south coast of Nova Scotia (Fig. 1) and is accessible by paved road. Fig. 1 provides directions from highway 103 and signs for "The Ovens Natural Park" are located along this route. From Highway 103 take Exit 11 and travel south toward Lunenburg; go approximately 8 km and turn right on Route 332; go approximately 12 km and turn left at Feltzen South turnoff; go approximately 3 km and turn right onto Ovens Park Road and follow 1 km to park.

Ovens Park and Historical Interest

The Ovens Park operates as a privately owned "natural park" which provides campsites, cottages, general store and a dinning room. The park features a walking trail and boat tours of several seacaves as well as information and activities related to the "Ovens gold rush", including, a walking tour, gold panning and a small museum. There is a small admission charge for a day pass into the park.

Gold Rush of 1861: Gold was first discovered in quartz veins in the cliffs in July of 1861 and in the beach sands in August of the same year. The ensuing gold rush resulted in a boom town of over a thousand miners. The placer deposits were quickly exhausted, principally under the direction of William Cunard, for whom Cunard Cove (Fig. 2) was named. No significant mining of the veins occurred and all efforts ceased within a couple of years. A small museum documenting the gold rush is maintained in the park and panning on the beach can still be rewarding for those seeking specimens.

The Ovens: The Ovens Natural Park was opened in 1935 with the principal attraction being several caves (Ovens) eroded into the cliff of the peninsula (Fig. 2). These caves can be observed from a walking trail above the cliff and, during the summer months, by boat trips. Although the caves formed mainly by natural erosion, Tuckers Tunnel was extended by mining activities. Local folklore states that a Mi'kmaq native once entered one of the caves by canoe and emerged on the Tusket River on the other side of the province.

Geological overview of Ovens

The area of the Ovens occurs within the Halifax Formation (Cunard Member) of the Meguma Group, the dominant unit of southern Nova Scotia. At the Ovens the strata consist mainly of dark slate with interbedded metasandstone (see D'Orsay 1980; Hall 1981 and O'Brien 1988 for further discussion of stratigraphy/sedimentology). The most prominent geological feature of the area is the Ovens Anticline, with coastal exposure of the hinge within the park and along the coast of Rose Bay (Fig. 2). The fold defines a classic chevron form with an interlimb angle of less than 40° (inset; Fig 2). The fold is slightly inclined to the north, plunges moderately to the northeast and slaty cleavage is typically axial planar (Fig. 2). Structures typical of chevron fold development are well developed, with clear evidence for flexural-slip folding late in the folding history; flexural slip refers to slip along bedding concordant horizons during folding (Fig. 3). In addition, abundant auriferous quartz veins are concentrated within the hinge area, including bedding-concordant and discordant veins. Two types of bedding-parallel veins are recognized: (1) Early bedding-parallel veins are characterized by tight buckling. This buckling is interpreted to have occurred prior to, or in the early stages of, folding. Fold development resulted in the parasitic asymmetry of buckled veins (Fig. 3). (2) Planar bedding-parallel veins occur along movement horizons formed during flexural slip folding. These veins commonly postdate discordant veins offset during flexural slip (Fig. 3). Although some veins were emplaced early in the deformation history, the majority of veins were emplaced along structures related to (flexural-slip) folding in the late stages of fold amplification. Fig. 3 is a schematic diagram of the Ovens Anticline illustrating folding-related structures and veins. (see O'Brien 1988; Horne and Clushaw 1994; and Henderson et al. 1992 for further discussion of structure and veining).

The coastal exposure in the Ovens area provides a good location for observing structure and veins in the Meguma Group and is a regular site for field trips by Dalhousie and St. Mary's universities and by local schools. Gold has been observed on all vein types discussed here, so keep your eyes open.

Field Guide

Following is a description of features which can be observed at various locations within the area. Locations 1–3 occur within the boundary of the Ovens Natural Park whereas locations 4–7 occur along the coast of Rose Bay (Fig 2). The main features of the Ovens Anticline can be seen within the park. The terrain in the area west of the park is somewhat treacherous and not recommended for large groups, such as school groups. However, this area should not be missed for those interested in fold structures and related veining and mineralization. Access to the Rose Bay section can be made in the following two ways: (1) walk along the coast from the park; (2) park your vehicle at the sharp bend in the road approximately 300 m before the park entrance (Fig. 2); a dirt road heading to the west will provide access to the coast. The land here is privately owned and you should ask permission from the residents. If walking from the park you will pass a drumlin exposed by erosion (Fig. 2). The drumlin is typical of drumlins in Nova Scotia, consisting mainly of red clay and well rounded boulders. Notice the abundance of florigen boulders (many of granite) on the beach.

Park Area

Location 1

Enter the park. Take the steps down to the beach from the parking area and walk to the bedding-parallel (NE-trending) cliff at the northern end of the beach. This is the beach where placer gold was mined. The hinge of the fold is exposed at the eastern end of the exposure (at the point where it becomes difficult to go any further), at which point you can stand on the hinge. Looking southwest (toward the park) the chevron shape of the fold can be observed. Bedding-parallel movement horizons reflecting flexural-slip folding can be seen on the north limb of the fold, some offsetting conjugate cross veins. Tightly buckled, early bedding-parallel veins (Fig. 3) folded at several scales straddle the hinge and reflect veins emplaced early in the deformational history. Note also the abundance of discordant veins (conjugate cross veins, Fig. 3).

Location 2

Walking trails lead along the cliff north of the parking lot. The caves (Ovens) eroded into the cliff can be seen from along this trail. Views from along this trail give an appreciation of the stratigraphy, consisting if slate with variable amounts of metasandstone. The first cave is referred to as Tuckers Tunnel and a set of stairs lead into the cave. Flexural slip bedding-parallel veins (Fig. 3) can be seen in the floor of the cave near the cave opening. Offset of the discordant conjugate cross veins at flexural slip veins supports the interpretation that these veins formed along flexural-slip horizons. From the trail above Tuckers Tunnel numerous flexural-slip horizons can be identified by offset cross veins on the north side of the inlet of the tunnel. The remainder of the trail is worth the walk for the view alone and look for good sedimentary erosional features on the bottom of metasandstone beds.

Location 3

Walk southeast of Cunard Cove along the shore. Note that the thick metasandstone beds diminish quickly, with stratigraphy dominated by slate and thin (10 cm) metasandstone. Steeply plunging, northwest trending kink folds are common in the slates. Striations and groves resulting from glaciation are common on flat, polished outcrop. Several roughly bedding-concordant mafic dikes are exposed on both shore of the Ovens peninsula (Fig 2). The age of the dikes is unknown, however, inclusions of cleaved slate indicates they post-date deformation (see Hall 1981 for more on dikes).

Rose Bay Section

Location 4

The first few hundred meters of exposure encountered along the Rose Bay section is strike-parallel, trending NNE, with cliff faces defined by bedding surfaces. Things to note include: abundant intersecting, conjugate cross veins (see Fig 3); local flexural-slip movement horizons, some with flexural slip bedding-concordant veins; sand volcanoes on upper surfaces of sandstone beds.

Location 5

This location includes the section from the end of the strike-parallel section of Location 4 up to the hinge zone exposed at Location 6. This section is slightly oblique to bedding, providing a transect through a portion of the south limb. The first part of this section is characterized by a series of strike-parallel thrusts with several slightly discordant ramps exposed. Bedding and cleavage dips have been rotated by thrusts and thrust-related folding, producing a complex profile section in this part of the limb. In addition, cleavage is typically not axial planar (steep) as in other parts of the fold and bedding-cleavage angles are atypically large, ranging up to 90 degrees. The large bedding-cleavage angles are interpreted to indicate that flexural shear strain related to folding was localized along the thrust planes, therefore preserving the prefolding cleavage-bedding relationship. Along the western end of this section bedding and cleavage are typical; both are steep and the bedding-cleavage angle is small. Flexural slip bedding-parallel veins are abundant along here right up to the hinge zone (Location 6).

Location 6

This is the point where the exposure becomes roughly strike parallel again. A vertical exposure of the hinge zone of the fold is exposed here; the cliff exposure to the west represents the north limb. Several features to note here. The chevron fold shape is evident looking toward the northeast. A hinge thrust (Fig. 3) transporting the north limb over the south limb is exposed at eye level on the north limb. This structure can be traced to the west along the same stratigraphic horizon to the next point (approx. 150 m). Anomalous amounts of quartz veins are exposed, including abundant cross veins and flexural slip bedding-parallel veins. Several bedding-parallel veins defining the hinge of the anticline can be observed in a horizontal exposure in the intertidal zone approximately 10 metres to the west (only visible at low tide). Coarse scheelite occurs in one of the veins. Note the local high concentration of arsenopyrite, typically within sandstone beds. Arsenopyrite is presumably related to veining.

Location 7

This is situated at the west side of a small point west of Location 6 (Fig. 2) where the hinge is once again exposed. It is only easily accessible at low tide. Note the abundance of bedding parallel and conjugate cross veins in the exposure between Locations 6 and 7. On the west side of the point, at Location 7, you can observe a vertical exposure through the hinge. A classic saddle reef vein occurs just above eye level and sandstone beds define a bulbose hinge structure characteristic of variable bed thickness.

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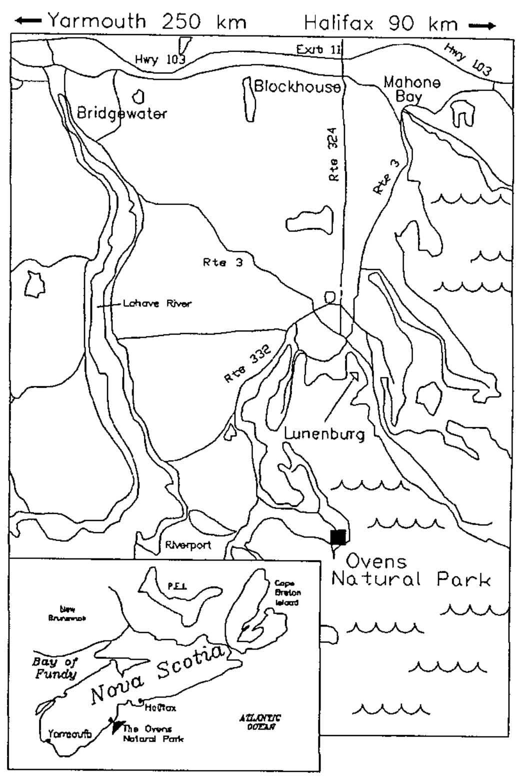

Figure 1: Location of the Ovens Park

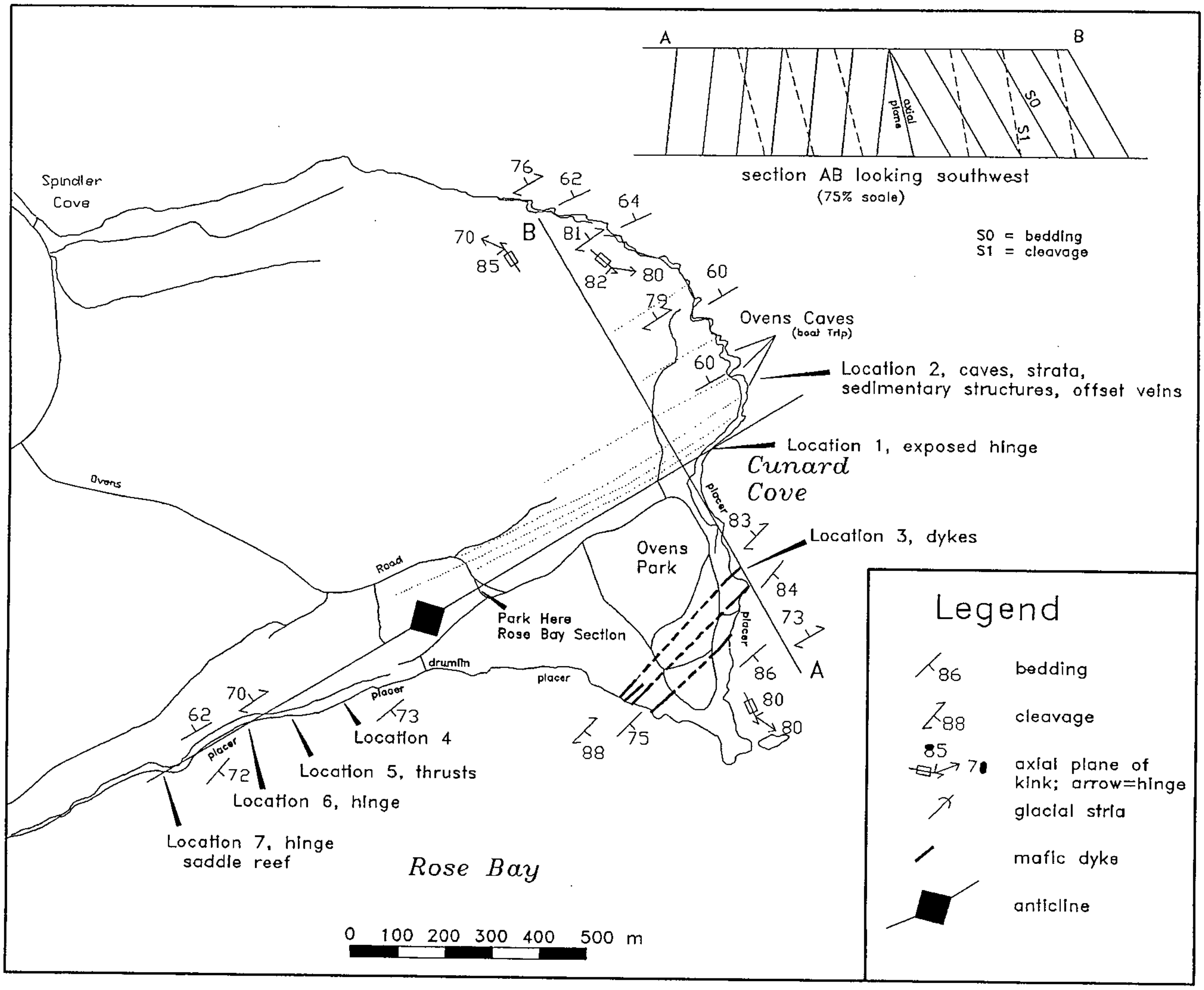

Figure 2: Geology map of the Ovens area showing locations discussed in text.

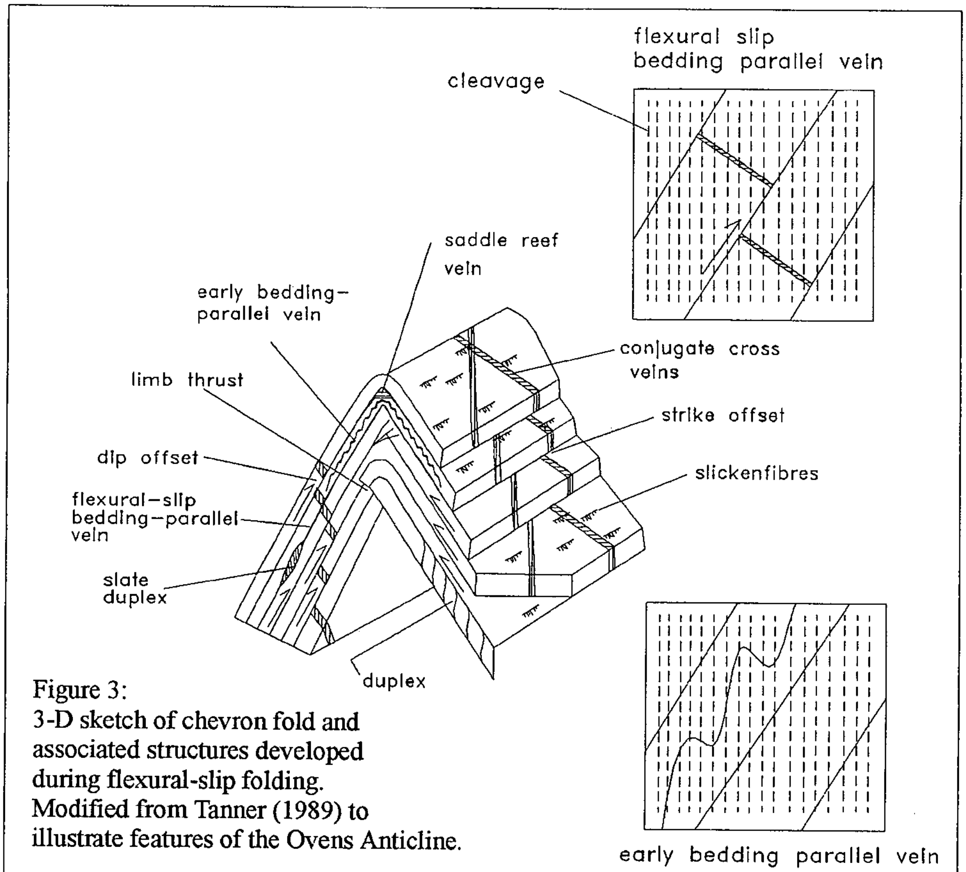

Figure 3:
3-D sketch of chevron fold and associated structures developed during flexural-slip folding. Modified from Tanner (1989) to illustrate features of the Ovens Anticline.

Big Tancook Island

Transition Rocks of the Goldenville-Halifax Formations

Location

Big Tancook Island is at the mouth of Mahone Bay and can be reached by passenger ferry from Chester, Nova Scotia.The ferry leaves the government wharf at various times throughout the weekdays and weekend, so consulting a schedule is important. The ferry is run by the Provincial Department of Transportation out of Blandford, phone (902)228-2340, and there is a fee.

Precautions

There are very few amenities on the Island, so bringing along snacks and liquids is adviseable. the weather can be variable. The ferry ride varies from windy and cool to hot, if you are out of the wind. The shore exposures can be wet, hot or cold, and in the evening the ferry ride back is generally cool to outright cold. Therefore bring layers of clothing that you can shed or put back on as the day progresses. Walking along the shore involves scrambling over ledges, so proper footwear is a must.

The island is private property, so respecting the rights of others is a foregone conclusion. do not sample, hammer, or otherwise deface outcrops, except where there are no houses and in the intertidal zone. Pay particular attention to your personal habits; pick up after yourselves and others.

Introduction

The Cambrian-Ordovician Meguma Group is a thick sequence of metamorphosed sedimentary rocks that underlies most of southern mainland Nova Scotia. The Meguma Group is quite unlike the rocks of equivalent age in northern Nova Scotia, New Brunswick, and Newfoundland, and it is thought that it was only brought into contact with the rest of what is now Atlantic Canada by movements along a major fault (the Glooscap Fault), which was a boundary between plates in late Paleozoic time. Southern Nova Scotia is distinguished as the "Meguma Terrane," the southernmost of the tectonic blocks that make up the Canadian Appalachians.

The Meguma Group has traditionally been divided into two formations. The lower of the two, the Goldenville Formation consists dominantly of metamorphosed sandstone, with subordinate slate, whereas the overlying Halifax Formation consists mainly of slate, with subordinate metamorphosed sandstone. The sandstones formations are interpreted as turbidite deposits—they were laid down by turbidity currents. A turbidity current is a variety of sediment gravity flow—a mass of water and suspended sediment that moved down a submarine slope into deep water. Turbidites are characteristically graded sandstone beds—they have sharp bases sometimes marked by erosional scours such as flutes and grooves, but their tops are gradational into the overlying fine-grained sediments, and show structures such as ripple cross-lamination indicating rapid deposition under the influence of a current.

Throughout most of the Meguma Terrane, the contact between the two formations is easily mappable, and younging directions close to the contact confirm that Halifax overlies Goldenville with a simple upward transition into finer-grained strata. However mapping by O'Brien and others (1986 a, b; O'Brien 1988) in the LaHave and Mahone Bay areas has shown that the transition in these areas is more complex. The basal part of the Halifax Formation consists of greenish slates and argillites that are highly enriched in manganese and other trace metals (e.g. Graves & Zentilli 1988). In places, a varied but mappable unit (the Tancook Member), including thinly bedded classical turbidites, occurs at the top of the Goldenville Formation. Elsewhere, a mappable sandstone-dominated unit (the West Dublin Member) is both overlain and underlain by slaty units. This complexity led to inconsistencies in the previous mapping of the Goldenville-Halifax boundary.

The legend on the attached map summarizes the inferred lithostratigraphy of the transition. One proposal has been to include all five transition zone members in a new formation (the Green Bay Formation). The map takes a more conservative approach, placing the Goldenville-Halifax boundary within the transition zone at the base of the greenish manganiferous sediments of the Mosher's Island Member. The overall succession of sedimentary facies in the transition zone is consistent with external control of sedimentation resulting from relative sea-level change in the source area from which turbidity currents were derived (Schenk 1991; Waldron 1987, 1992).

The Meguma strata have been folded in a major regional deformation dated around the Early Devonian (about 400 Ma), and the fine grained sedimentary rocks are generally strongly cleaved. Metamorphism at low greenschist facies has partially obliterated sedimentary textures in thin section, but sedimentary structures are generally clear in outcrop. Further metamorphism resulted from the intrusion of the South Mountain Bathoilith around 370 Ma.

Big Tancook Island

The east coast of Big Tancook Island exposes the most complete section of the Goldenville-Halifax Transition. A laterally equivalent partial section is exposed on little Tancook Island, but the ferry schedule does not conveniently permit visits to both in the same day. Ferry times are posted prominently on the wharf at Chester, and are also obtainable from the Nova Scotia Department of Tourism. The ferry is for foot passengers only. Fortunately much of the coastal section through the lower Halifax and upper Goldenville Formation is accessible in a day's walk. The shore is locally very bouldery, making walking tiresome; for most of the way there is a path in the trees just inland from the rocky beach. The section will be described from top to bottom, this being the order in which the units are encountered on a walk starting at the ferry wharf (Fig. 1). More detailed geological descriptions may be found in Waldron 1987, 1992.

Area 1: At the north extremity of the island, the highest part of the section is seen. These are black slates and pyrite-bearing siltstones of the Cunard Member. The siltstone layers are very thin (mostly a few millimetres to a few centimetres), but there are a few thicker beds of massive grey very fine sandstone. Some beds are lenticular, with rather large-scale cross-laminations. There is a strong cleavage, oblique to bedding, and with a somewhat steeper dip. Cubic pyrite crystals up to a centimetre across can be found in some of the siltstone beds.

Also in this area there are cliffs of poorly consolidated glacial till, from which you may be able to recover pieces of fossiliferous Windsor Group (Carboniferous) limestone. Outcrop of Windsor Group has been found at various locations around the coast of Mahone Bay, but none has been recorded on the Tancook Islands (Fig. 2).

As you walk east along the shore, you are crossing obliquely onto older strata. Down section, there is a transition into generally grey-green manganese-rich argillites of the Mosher's Island Member, though this transition is incompletely exposed in outcrops that are close to the low-water mark.

Area 2: The Mosher's Island Member is the most laterally continuous of the units in the Goldenville-Halifax transition, being traced throughout the area mapped by O'Brien et al. These metamorphosed fine-grained sediments are sometimes termed argillites because they are less well cleaved than the overlying slates. Locally they contain concretions of manganese-bearing calcite and spessartine garnet which weather black. The argillites are mostly laminated and show only very small-scale ripple cross-laminations and occasional structures produced both by soft-sediment deformation. They represent deep-water sedimentation under low-energy conditions by mud turbidites or other bottom-following currents.

Area 3: The top of the Tancook Member is marked by the appearance of sandstone beds with ripple cross-laminations. In the uppermost part of the Tancook member (unit 4), just south of Hutt's Point, sandstones are scarce. Slates are interbedded with thin beds of manganoan carbonate. Locally, the more rigid carbonate beds are folded in small scale asymmetric folds. The cleavage is axial planar to these folds. Elsewhere, the cleavage is deformed by later kink folds.

Area 4: Lower in unit 4 of the Tancook member, sandstones occur grouped together in sporadic 'packets' which can be correlated with ease across the approximately 1 km interval between Big and Little Tancook Islands. Note that the uppermost part of the Member seems to lack bioturbation structures, but abundant burrows of *Teichichnus* type appear abruptly at a package of sandstone beds informally named the 'outhouse beds' because of their uniquely picturesque location here. This packet can be correlated throughout the Tancook Islands and onto the adjacent mainland.

Area 5: Down section, sandstone packets become more common, marking the transition to unit 3 of the Tancook member. Large concretions of Manganoan calcite are present in some of the sandstones; these are preferentially weathered to produce depressions in the outcrop surface that are locally known as "devil's footprints." Close to the point where sandstone becomes dominant, a unique 45 cm bed of shell debris is encountered. The bed is deeply weathered to dark brown (because of its manganese content) but is grey on fresh surfaces. Early Middle Cambrian trilobite fragments have been identified in this bed (Pratt & Waldron 1991). Please note that this is a unique outcrop in the Meguma Terrane and the amount of material is small; the fossil material is not of itself spectacular in appearance; please do not collect unless you will do serious research and publish the results!.

Area 6: About in the middle of the Tancook member is a unit (termed unit 2 in the stratigraphic columns) totally devoid of sandstones. Note that the manganiferous argillites are mostly unlaminated, showing a 'mottled' appearance typical of heavily burrowed muds. In places, probable burrows are preserved in carbonate or pyrite. The unit begins abruptly above a thick but heavily bioturbated sandstone. The most likely hypothesis for the origin of this bed is that a relative sea-level change or other event in the source area temporarily cut off the supply of sand-laden turbidity currents to the area. Without the repeated rapid depositional events, a fauna of burrowing organisms was able to flourish. The bioturbation (disturbance of the sediment by organisms), both here and in the rest of the Tancook Member, shows that the sea-floor was oxygenated. In contrast, the black slates of the Cunard Member (area 1) were probably deposited on a sea-floor that was depleted in oxygen.

Area 7) The lower part of the Tancook Member again is mostly sandstone, but the sands are thinner bedded and are less clumped into packets of thin and thick beds. These resemble deposits found in large 'lobes' of sand in the outer parts of modern submarine fans. Many of these beds show grading, sharp bases with flute and load structures, laminations, and cross-laminations. The structures are organized in Bouma sequences characteristic of turbidity current deposits. Interestingly, the current-generated structures, both here and in the rest of the Mahone Bay area, show that most of the sand was deposited by currents that flowed north or northeast, apparently from the direction of the present-day Atlantic Ocean. Of course, the Meguma Group is much older than the opening of the Atlantic Ocean; the sand was probably transported from an adjacent area of the ancient continent of Gondwanaland—comprising modern Africa, South America, India, Australia, and Antarctica—which lay in the area to the southeast of the modern Meguma Terrane. To the southwest, at flat reef, bedding is almost horizontal on the crest of a structural dome on the hinge of the Indian Path Anticline.

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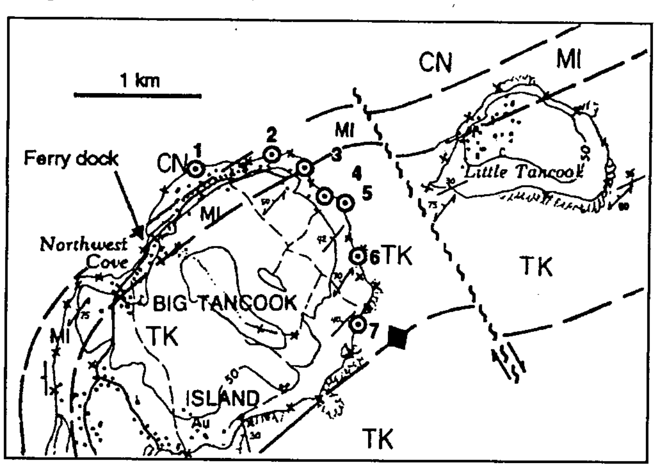

Figure 1: Location map for stops on Big Tancoook Island. Map also includes geology from O'Brien 1986; TK - Tancook Member, MI - Mosher Island member, CN Cunard Member.

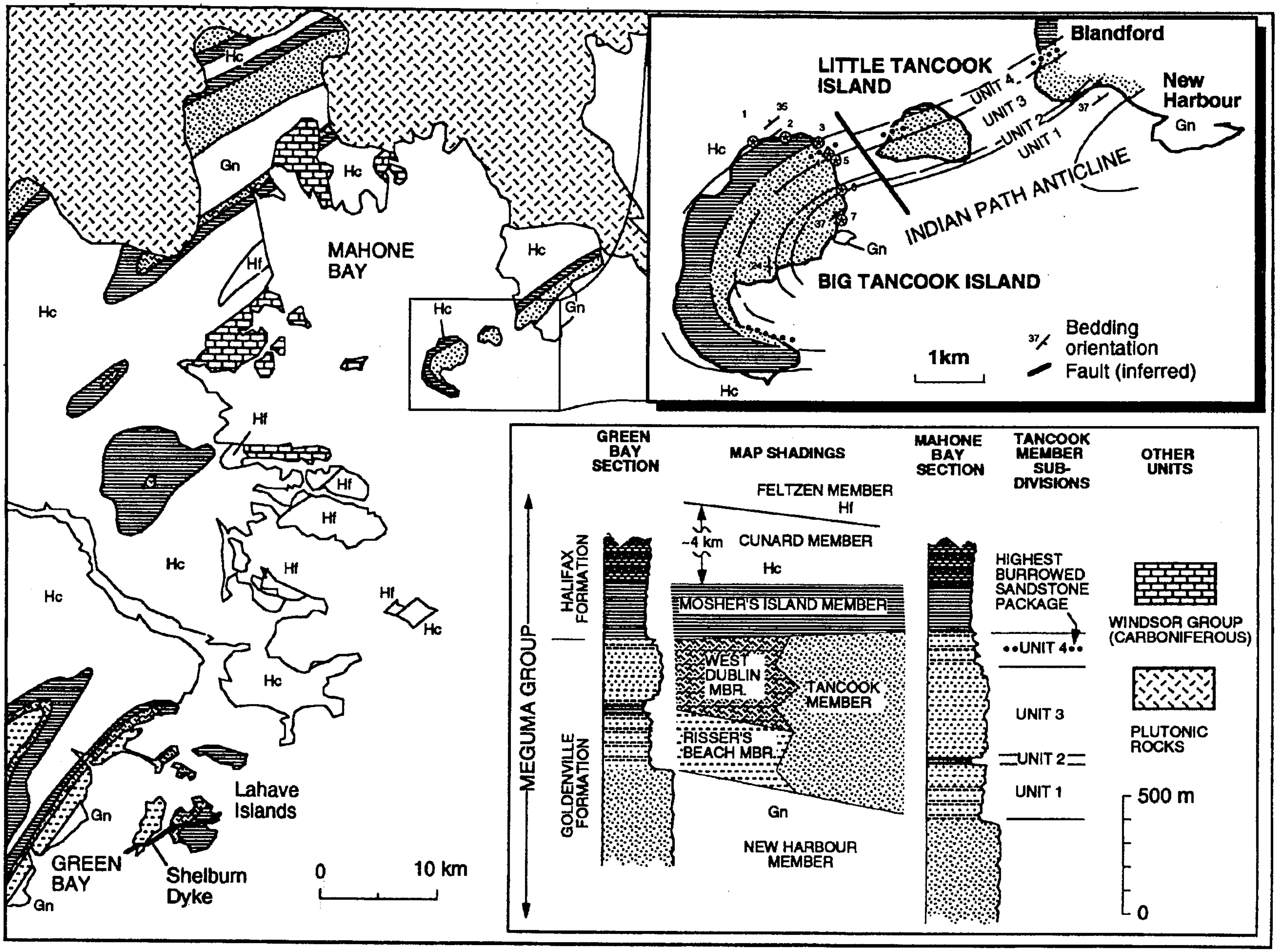

Figure2: Detailed geology maps and sections for Mahone Bay area.

Peggys Cove

A Piece of Nova Scotia

Introduction

For years, the picturesque combination of ocean and rugged granite coastline of Peggys Cove has provided artists and travellers with a setting of impressive natural beauty. The observable landforms and bedrock features provide geologists with valuable information about the geological history of Peggys Cove that stretches over 470 million years.

Caution

Although many people enjoy gazing at the panoramic view of breaking waves at Peggys Cove, we urge you to exercise extreme caution. Large waves often break further on shore than expected. Visitors can "play it safe" by avoiding shoreline edges and steep slopes. Please ask permission to cross private property.

Geological History

The geological history of the Peggys Cove area began over 470 million years ago not with granite but mud and sand. These sediments were being deposited in a deep ocean basin at the south pole near ancient Africa. As the thickness of sediment in increased, the mud and sand were compacted into rocks named shale and sandstone. As thousands of metres of sediments accumulated, the entire ocean basin and adjacent continents were moving toward ancient North America. The collision of Africa (really Gondwanaland) and North America initiated a period of mountain building 390 million years ago. The shale and sandstone were crumpled and folded, and the added heat and pressure transformed them into metamorphic rocks known as slate and quartzite.

The geological history of the Peggys Cove area began over 470 million years ago. Although granite forms the surface rocks now, it was not always present. Mud and sand were the originally deposited in a deep ocean basin near the south pole adjacent to a huge continent known as Gondwana. Slowly the deposits accumulated and were compacted into shale and sandstone.

Movement of the ocean basin and Gondwana toward ancient North America eventually resulted in a major collision of crustal plates. This collision initiated a period of mountain building 390 million years ago that formed part of the Appalachian Mountain belt. The shale and sandstone were crumpled and folded, and the added heat and pressure transformed them into metamorphic rocks known as slate and quartzite.

Heat generated from the collision melted rocks at the base of the earth's crust somewhere below 40 km. The molten material intruded upward because it was less dense and finally crystallized into granite at 370 miiljon years ago. Over the next 350 million years uplift, weathering and erosion removed many kilometres of quartzite, slate and granite. The granite visible today was once buried deeply in the ancient mountain belt.

About 350 million years ago when the climate was tropical and the water warm, the sea invaded all the valleys. Conditions were perfect for the building of reefs and the precipitation of limestone, salt, gypsum, and potash. One of the low-lying valleys invaded by the sea is now St. Margarets Bay. Limestone deposits have been found on the floor of the Bay and on its east side, south of Tantallon. This limestone represents the deposits of a warm, inland sea.

Today, we know the states and quartzites as the gold-bearing Meguma Group. The many gold mines in Nova Scotia during the period 1865-1950 were located in rocks of this Group. The granite is now referred to as the South Mountain Batholith which extends from Halifax to Yarmouth. Near Yarmouth the granite holds the East Kemptville tin deposit. Many quarries have been developed in the granite for building stone.

Granite Forms the Bedrock

Peggys Cove and surrounding areas are underlain by coarse grained, greyish-white granite. Granite is a hard igneous rock formed by the slow cooling of hot molten material into coarse-grained crystals of quartz, feldspar (orthoclase and plagioclase), and mica (muscovite and biotite).

An interesting geological feature in the granite is the presence of dark-coloured fragments known as xenoliths. As the hot, fluid granite pushed its way upward, fragments of layered bedrock sometimes fell into the granite melt and were almost digested by the hot magma. Other xenoliths represent fragments of an earlier solidified granite. Either type of xenoliths appears as dark inclusions both layered and unlayered up to 20 cm in size.

Cutting across the granite are veins of younger aplite and peginatite. These two rock types were once hot and fluid; they were intruded into the granite after it solidified. Pegmatite can be distinguished from aplite because it has much larger crystals (1-6 cm), indicating that the rock cooled very slowly. Aplite is finer grained and has a sugary texture. The smaller crystal size (1-2 mm) indicates a faster cooling process.

Fractures and Faults

Commonly, the granite of this area is fractured and faulted with oriented cracks that formed during the cooling of the granite and mountain building. Two prominent directions of fracturing are shown in the granite: one is north-south and the other is northwest.

Glaciers Sculpted the Landforms

A huge ice sheet, several kilometres thick, covered much of Nova Scotia between 10 000 and 70 000 years aco. Glacial action yielded distinctive erosional and depositional features.

Erosion of the landscape was accomplished by the gouging and plucking of granite bedrock along zones of weakness such as fractures. Roches moutonnées are mounds of rock which have been named because of their resemblance to reclining sheep. Such a mound displays a gradual smooth slope on the north side from which the glacier came and a rough steep slope on the other side. During glaciation, the ice used the weakness of the fractures and faults in this area by simultaneously widening them and removing rock material by "plucking." Peggys Cove owes its dog leg shape to glacial erosion of north and northwest trending fractures.

Rocks at the base of moving glaciers left scratches called glacial striations on the bedrock surface. Bogs and small marshy lakes are also features of glaciation; they formed because of poor drainage of the hollows that were eroded by glacial action.

As they moved, the glaciers also deposited a mixture of unsorted rocks, gravel, sand, and clay known as glacial till. Deposition formed oval shaped hills known as drumlins. Familiar examples in the Halifax area are Citadel Hill, hills on McNabs Island, and Georges Island. There is also a series of drumlins along the east coast if St. Margarets Bay. As the glaciers retreated, boulders or erratics were deposited. Today, numerous erratics can be seen on the landscape of Peggys Cove.

When the ice sheets melted, sea level rose and filled these coastal areas. Today, the mechanical action of breaking waves and repetitive freezing and thawing continues the process of erosion by removing chunks of granite from the edges of coast.

Field Trip Stops

Each of these stops is numbered on the map.

STOP 1. William E. degarthe Memorial Provincial Park. The granite ledge was carved by the late artist William E. deGarthe as a monument to the fishermen of the region. Granite is an excellent medium for carving because of its extreme durability. However, costs of working granite are high because it is very hard.

STOP 2. Peggys Cove. The Cove results from glacial erosion along a-two sets of closely spaced fractures. The northwest and north trends account for the dog-leg shape of the cove.

STOP 3. Lighthouse. Granite is composed of the minerals quartz, orthoclase and plagioclase feldspars, and biotite and muscovite micas. The large crystal size indicates slow cooling.

STOP 4. Lighthouse. Abundant fragments of preexisting rocks (xenoliths) are common in the granite. They represent small and large blocks of the underlying rocks that were tom off by the intruding granite from preexisting rock. Both layered and non-layered xenoliths can be seen here.

STOP 5. On the rocks cast of the Lighthouse. Visible here are large roches moutonnées or "sheep backs," formed by glacial erosion. Notice that the steep sides are south facing indicating glacial ice movement toward the south. Rocks lodged in the base of the glacier left scratches (glacial striations) on the granite. From this position one can see many large and small glacial erratics on the northern skyline.

Dykes and veins cut the granite in great abundance here. Many are made up of a coarse-grained rock called pegmatite which contains large crystals of quartz, orthoclase and plagioclase feldspars, tourmaline, and muscovite. Some veins are composed of a finer grained rock called aplite. The veins of aplite and pegmatite are younger than the granite.

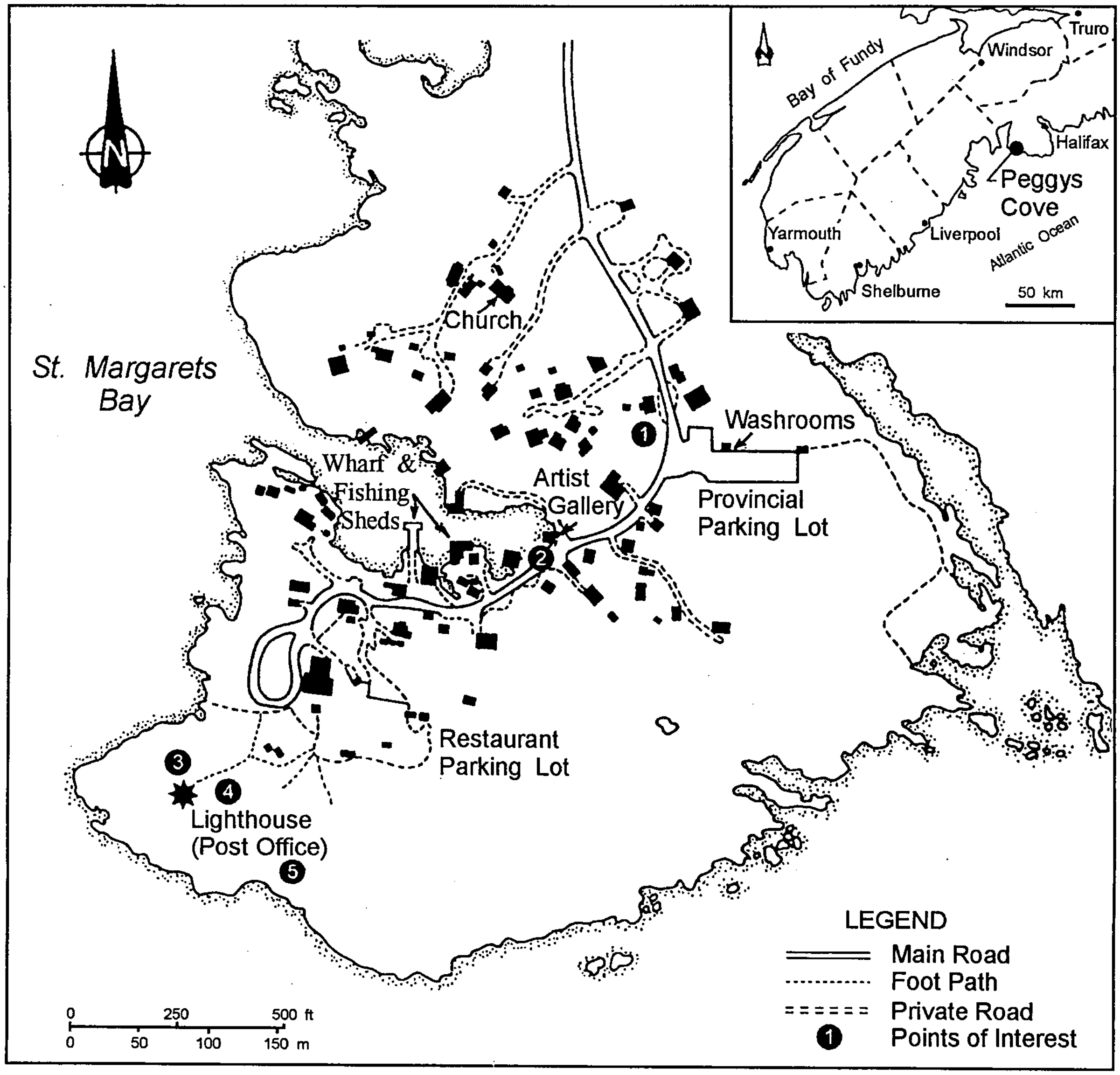

Figure 1: Location map of Peggys Cove showing points of interest described

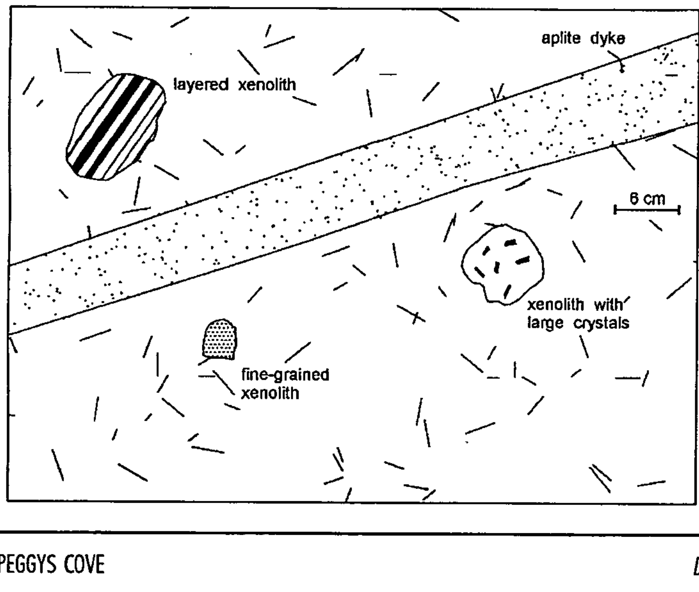

Figure 2: Sketch showing typical structures found in the rocks at Peggys Cove

Chebucto Head

Layered Granites and Cross Cutting Dikes

Purpose/Highlights

- foliated porphyritic granodiorite
- xenoliths and inclusions
- layered granodiorite
- aplite, pegmatite and aplite/pegmatite veins and dikes

Location

Take Route 349 (Herring Cove Road) out of Halifax to Duncans Cove. Turn off to Duncans Cove and continue the road to Chebucto Head (Fig. 1). Park either by the lighthouse or at one of the many parking areas along the road. Walk down to the rocks at the shore.

Special Precautions

This trip is in an area where there are wet and slippery rocks, steep drops, and energetic, unpredictable waves. Please exercise extreme caution and watch out for others. A good safety practice is to examine the rocks with others and not to wander off alone.

You will be in an area of large boulders, steep slopes, and deep chasms. It is best not to jump from boulder to boulder, climb up the steep slopes, nor to jump across the chasms; go around and have one foot on the ground at all times. This is an area for good sturdy field boots with soles for gripping rock. Do not take the ocean for granted.

Introduction

The Halifax metro area is underlain by the slates and quartzites of the Meguma Group of Cambro-Ordovician age which were folded about east-west axes during the Acadian Orogeny (400-375 Ma). These were then intruded by the South Mountain Batholith about 370 Ma (372-361 Ma, Clarke and Halliday 1980; 367 ± 4 Ma, Reynolds et al. 1981; plus others). In this area the contacts of the granite with the country rock are sharp and discordant, truncating regional trends. Along the western approaches to Halifax Harbour the granite passively stoped the country rock, the best examples being around Portuguese Cove but seen along the whole margin on the Batholith and evident in the abundant xenoliths in the granite itself.

The South Mountain Batholith has been mapped into several lithologies on the basis of mineral proportions (MacDonald and Horne 1987). All are mainly massive and unfoliated but some areas along the margins and near the coast show a weak to strong foliation fabric. Some sheeted quartz-greisen veins and dikes, along with aplite and aplite/pegmatite dikes, subparallel the contact with the country rocks and many areas are cut by later (post Acadian deformation) localized shear zones and faults.

A fabric of subparallel alignment of alkali feldspar megacrysts and inclusions is evident in many outcrops. It can be both pervasive and linear across a whole outcrop or swirl about over several metres. In many instances these swirls are associated with larger xenoliths or inclusions, especially in the stoped regions. Across the whole area the foliations seem to define dome-like structures which probably were related to movements in the magma during the final stages of emplacement and crystallization.

Layering and schlieren are seen occasionally in outcrops along the coast and at the margins of the batholith. The best exposed layers are at Chebucto Head. The layers are on the order of 20-30 cm thick and appear very much like sedimentary structures. The schlieren, where seen, are different from the layers in that they have diffuse boundaries and are small. Both are different and distinct from the fabric defined by the phenocrysts and inclusions.

Description

The South Mountain Batholith at this locality is generally a homogeneous porphyritic light- to medium-grey biotite granodiorite. It is coarse-grained with approximately 10-15 per cent euhedral alkali feldspar megacrysts between 2.5 and 7 cm in length. The groundmass is composed of quartz, plagioclase, K-feldspar, and about 5-10 per cent biotite with a trace amount of muscovite. Throughout the outcrop area are rounded mafic inclusions and rounded to angular xenoliths with relict bedding. The mafic inclusions are generally porphyritic with plagioclase and hornblende phenocrysts. Many of the xenoliths are quite large, extending up to 80 cm or more across, and show both endocontact (inside the xenolith) as well as exocontact (outside the xenolith in the granodiorite) reactions.

Underneath the Chebucto Head lighthouse (Fig. 1), the granodiorite displays a restrictive area of layers dipping shallowly to the east. The layers are variable in composition and texture ranging from equigranular to varying amounts of small megacrysts up to 60 per cent or more to variable mineralogy from leucocratic to melanocratic (0-40 per cent mafics). The layers range in thickness from a few cm to 1 m (average about 30 cm) and are quite tabular and continuous across the outcrop area of several hundred metres. The upper contact or margin of the layered area with the main phase granodiorite is quite distinct and planar with minor undulations. In some places the contact is a pegmatite vein but the timing of the intrusion and its age relationship with the layering is not clear. In the granodiorite, to about 30 cm from the contact, there is a crude parallel alignment of megacrysts to the margin.

Indistinct and smaller areas of layering better defined as schlieren can be seen to the south towards the bunker ruins; many are just zones, accumulations, or layers of megacrysts. These schlieren are distinct from the foliation or alignment of megacrysts which is pervasive over the whole area.

In walking from the lighthouse to the area of the ruins you pass many thin (10–30 cm wide) but continuous (over 100 m long) pegmatite and aplite-pegmatite veins. Their mineralogy is quartz, K-feldspar, muscovite and tourmaline and many are zoned with quartz ± muscovite ± tourmaline along their central axes. Some veins branch and others show cross-cutting relationships with other veins. Some veins cut the schlieren or high megacryst accumulations. The vein and dike contacts vary from sharp to diffuse indicating variable temperature differences between rocks as well as possible different ages of dike intrusions.

Also in this walk you pass many areas of intense fracturing, late foliation, and hematization of the feldspars. In two such areas, about 200 and 400 m south of the lighthouse, the granite exhibits spheroidal weathering (spheres <1 m in diameter) within these zones. In the latter area a pegmatite vein crosses the zone but is unaffected by the foliation and hematization indicating either strong competence or a later age.

In the area of the bunker ruins near to the water's edge (Fig. 1) are two large (>1.5 m) aplite dikes oriented parallel to the coast and numerous randomly oriented small aplite, pegmatite, and aplite-pegmatite veins and dikes. Parts of the dikes are sheeted with alternating layers of pegmatite and aplite and/or show crude aplitic layered schlieren. Zones of extreme foliation occur throughout the dikes as well as along their margins and in the country rock to about 20 cm. Several of these foliation zones are consistently parallel to the boundaries of the large dikes but other orientations can be seen in the adjacent areas.

Further south from the bunker ruins is more layering, similar to that seen beneath the lighthouse. The textures, mineralogy, and layering follow the same pattern as seen to the east, but some outcrops are more three-dimensional so therefore more spectacular. Here some layers bifurcate into two or more branches, some layers have well developed scour structures resembling channels in sedimentary rocks. One bed shows a well developed cross-bedded structure defined by biotite and feldspar phenocrysts, and several layers are folded similar to soft-sediment slumps. Each layer is unique and distinct and the main zone of layering is sharply defined by later dikes with one major dike full of phenocyrsts. Outside the main area are several smaller zones of layered granodiorite with diffuse boundaries sub-parallel to the coast. The coarse fraction (phenocrysts and xenoliths) in this area is restricted to the granodiorite and the cross-cutting dikes, the layers are generally free of these coarse constituents except for the occasional band.

Conclusion

The origin of the layering in granites is a hotly debated issue and several models exist. There is no evidence of cataclasis in the mineral grains and the contacts between the layers and the main granodiorite pluton varies between sharp and distinct to gradational. Assimilation of inclusions grading into the layers cannot be seen anywhere so this mechanism is ruled out. Most layers seen in the SMB are close to the margins or contact with the country rock so most likely it formed in the last stages of pluton emplacement. Smith (1975) proposed differential shear during flow for the layers whereby velocity gradients sorted the very coarse and fine to coarse fractions. Recently Clarke and Clarke (in press) have refined the original proposed model of 1980 of repeated roof-block subsidence to account for the layers. In this model a congealed block of granodiorite near to the margin cooled and contracted which resulted in its subsidence into the main magma beneath it. The resulting fracture space was then invaded by magma from below through a dike system. This occurred repeatedly (modelled every 10–25 years) and each pulse is represented by a layer.

After the layering events the whole region was cut by late-stage aplite and aplite-pegmatite dikes. The granodiorite at the time of dike emplacement was most likely similar to a visco-elastic medium, breaking catastrophically during initial injection but then flowing and shearing during the waning stages along the margins with the dikes. The crystallization history of the dikes themselves is also probably similar to that of the granodiorite and resulted in their internal foliation.

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Source

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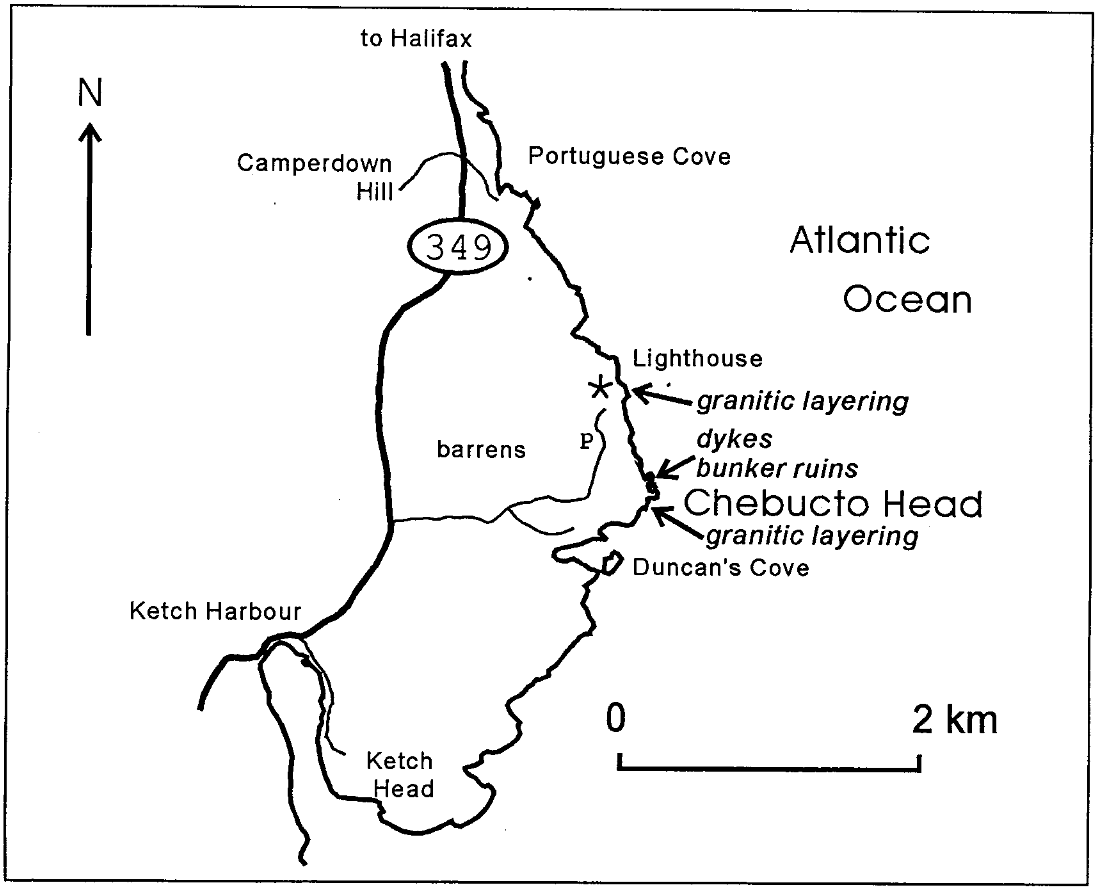

Figure 1: Location map with areas of geological interest outlined.

Bayers Lake Business Park

The Meguma Group-South Mountain Batholith Contact

Highlights

- intrusive contact of the South Mountain Batholith
- aplite and pegmatite veins and dikes
- tourmaline and garnet

Directions

The Bayers Lake Business Park is to the west of the Halifax Peninsula between Highways 102 and 103 (see Fig. 1). The first outcrop is off the 102 and the second one is off 103. Either outcrop can be viewed before the other but it makes more geological sense to view them in the order described.

For stop 1 take interchange 2A off the Bicentennial Highway and turn towards the Bayers Lake Business Park on Chain Lake Drive. After the sharp turn take a left into the Price Club Parking lot and park close to the interchange. Walk back down the drive to Chain Lake Drive and head towards the interchange.

To get to the second outcrop either go back on 102 and take the number 2 interchange to 103 or drive through the Park on Chain Lake Drive to the Bay Road or Route 333. Either way study the map (Fig. 1) because there are many twists and turns and one-way ramps. The only location to park is off the on-ramp onto 103 from Route 333 (Fig. 1) or the St. Margarets Bay Road. Restrict observations to the outcrops in the immediate area.

Special Considerations

These outcrops are at busy interchanges on major highways but are not on the highways themselves. DO NOT STEP ONTO THE ROAD BEDS. This means, stay next to the outcrops or in the ditch next to the outcrops. Do not go onto the gravel next to the pavement. Be especially careful at Stop 2 because you will be stopping with a vehicle next to an outcrop. Other travellers will think there has been an accident and will probably stop as well; maybe you should post a large lettered sign in your back window saying you are doing a geological survey.

Introduction

The Halifax metro area is underlain by the slates and quartzites of the Meguma Group of Cambro-Ordovician age which were folded about east-west axes during the Acadian Orogeny (400-375 Ma). These were then intruded by the South Mountain Batholith (SMB) about 370 Ma (372-361 Ma, Clarke and Halliday, 1980; 367 ± 4 Ma, Reynolds et al., 1981; plus others). In this area the contacts of the SMB with the country rock is sharp and discordant, truncating regional trends. In most areas the SMB passively stoped the country rock with the best examples being seen along the eastern coastal exposures of the Halifax Peninsula but is evident along the whole margin of the Batholith and in the abundant xenoliths in the granite complex itself.

Through bedrock mapping in the Halifax Peninsula, the SMB has been divided into several lithologies based on mineral proportions (MacDonald and Horne, 1987). All are massive and unfoliated with a few later (post Acadian deformation) localized shear zones and faults. Late quartz-greisen veins and dikes, and aplite and aplite/pegmatite dikes cut the granite in many localities.

A fabric of subparallel alignment of alkali feldspar megacrysts and inclusions is evident in many outcrops. It can be both pervasive and linear across a whole outcrop or swirl about over several metres. In many instances these swirls are associated with larger xenoliths or inclusions, especially in the stoped regions.

This field trip will examine the various types of igneous rocks exposed in the Bayers Lake Power Centre area and the relationships between those intrusives and the country rocks. The first outcrop shows the SMB (granodiorite) intruding the Meguma Group (here the country rock) and the second outcrop shows aplite and aplite/pegmatite dikes and veins intruding the SMB (a different country rock).

Stop 1—Interchange 2A, Bicentennial Highway 102

The South Mountain Batholith at this locality is a medium grey, medium to coarse-grained granodiorite. It is generally equigranular to weakly megacrystic. Biotite content ranges from 10 to 20 per cent, there are trace amounts of muscovite, and K-feldspar rarely exceeds 5 per cent. There are abundant xenoliths and inclusions of various sizes throughout the outcrop and in the boulders that make up part of the wall face. There are a few small pegmatitic veins but they are not prominent.

The country rock here is in the transition zone between the Halifax and Goldenville Formations of the Meguma Group. It is quite massive and quartz rich.

The texture of the Meguma is that of a hornfels, a medium-grained granulose rock with no cleavage, schistosity, or parallel alignment of minerals due to metamorphism. Relict bedding can be seen on some faces as colour laminations, dipping steeply to the southeast.

The contact between the granodiorite and the Meguma Group is quite distinct as a zone of mixing of the two rock types over about seven metres. Irregular pods of granodiorite with fine grained margins and coarser interiors invade the country rock and irregular pieces of the country rock can be found surrounded completely by granodiorite having been broken off and carried away by the magma (stoping process). The bedding planes of the Meguma trend right into the pluton, showing the cross-cutting relationship.

Stop 2—Interchange 2, Highway 103

The South Mountain Batholith at this locality is generally a homogeneous megacrystic light- to medium-grey biotite monzogranite. It is medium to coarse-grained with approximately 10 per cent subhedral to anhedral white alkali feldspar megacrysts <2 cm in length. It is composed of 10-15 per cent biotite, trace amounts of muscovite and cordierite, and equal amounts of quartz, plagioclase and K-feldspar. Throughout, the outcrop area is rounded to angular xenoliths with relict bedding and rounded mafic inclusions. Many of the xenoliths are quite large, extending up to 80 cm or more across, and show both endocontact as well as exocontact reactions with the granite. The mafic inclusions are generally porphyritic with plagioclase and hornblende phenocrysts.

Cutting the granite are many aplite/pegmatite dikes and veins, some up to 1 m across. Their mineralogy is mainly quartz, pink K-feldspar, and muscovite with traces of tourmaline, garnet, and cordierite. Textures range from aplitic to coarsely pegmatitic and graphic with gradations between the three. Compositional layering is marked in some dikes by central areas concentrated in quartz or muscovite or both, marginal zones of K-feldspar, and restriction of tourmaline and garnet to the central zones.

Discussion

Interpretation of the geology from these two stops shows the decreasing age relationships between the Meguma Group, the South Mountain Batholith, and the aplite/pegmatite veins and dikes. They also show the methods by which igneous rocks intrude country rocks. The main pluton intruded by either breaking off pieces of country rock and removing them from the immediate area while moving into the spaces previously occupied (known as stoping) or by melting parts of the magma chamber walls thereby incorporating part of the country rock into the magma itself. This results in irregular margins between the intrusives and the country rock. The dikes and veins, however, only intruded by forceful injection along fractures and fissures which results in tabular shaped bodies with sharp boundaries.

If a magma cools and crystallizes by giving off heat to the country rock then that part closest to the country rock will crystallize first. That portion in the middle of the magma chamber will crystallize later. During this process of solidification and crystallization inwards, two things happen. First, early formed minerals (those that crystallize at higher temperatures) are prevented from mixing and equilibrating with the residual parent liquid because they are locked into the newly formed rock at the margins (a process called fractional crystallization). This leads to a series of igneous rocks of different compositions from the margins inwards. The different igneous melts in the center of the magma chamber can then intrude outwards along cracks and fissures and ultimately find their way into the first formed solid rock along the margins. This accounts for the different mineralogies between the granite and the dikes even though they came from the same pluton. Second, some elements such as copper (Cu), tin (Sn), boron (B), water (H_2O), and carbon dioxide (CO_2), etc. do not go readily into the early formed minerals so they concentrate in the last pockets of magma and give rise to exotic minerals (such as tourmaline and garnet) in the late stage dike rocks.

According to another theory (Jahns and Burnham 1969) aplite is typically a quench texture resulting from the catastrophic pressure drop due to opening up of fissures or from a quickly ascending crystallizing magma. The volatiles (H_2O and CO_2) released by this process then allows pegmatites to form in the later crystallization sequence. The ratio of aplite to pegmatite gives the observer a rough measure of the amount of volatiles in the melt or what was concentrated in the residual melt after the main part of the granite crystallized. Here pegmatite volume is greater than aplite volume indicating a fair amount of volatiles.

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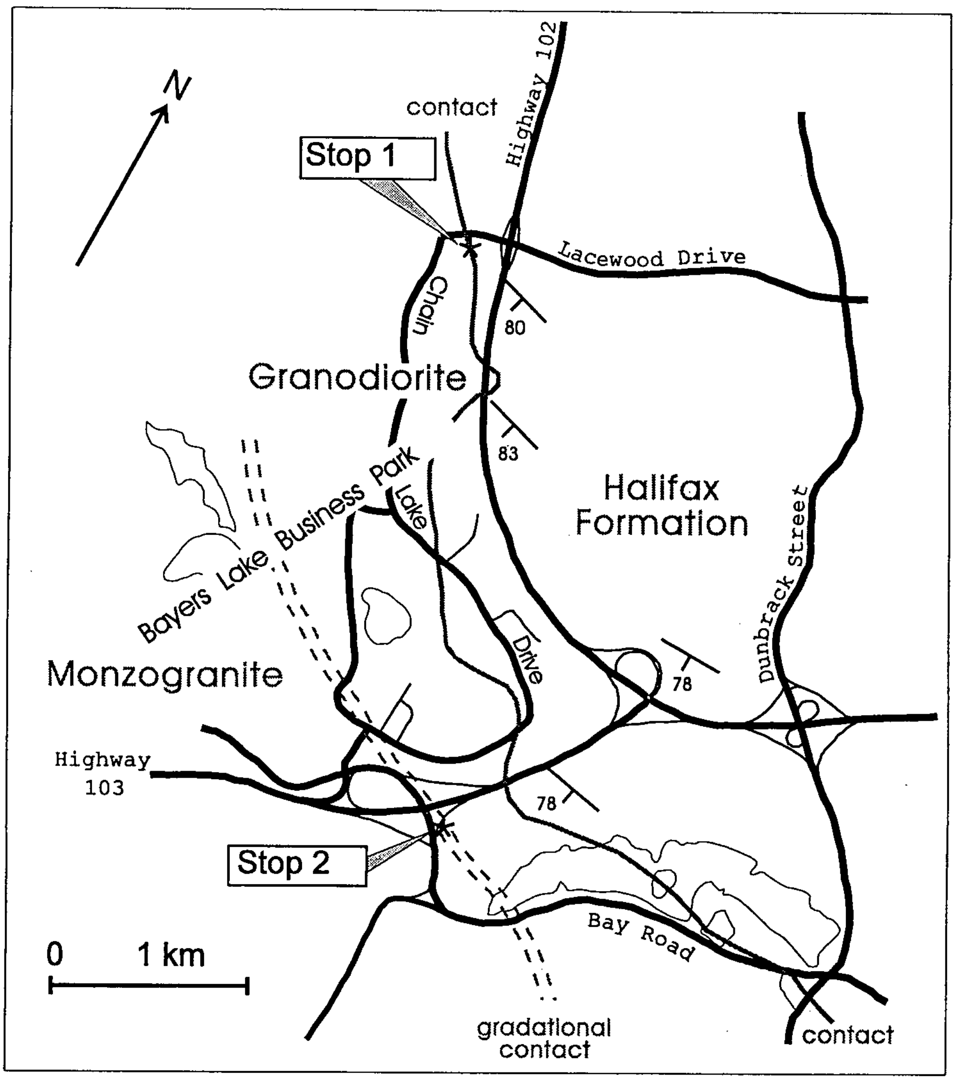

Figure 1: Location map showing geology of Bayers Lake Business Park. Geology from Macdonald and Horne, 1987.

Point Pleasant Park

Bedrock Erosion by Glacier Ice and Flowing Water

Highlights

- erosional features produced by glacier ice
- erosional features produced by running water

Directions

Point Pleasant Park is located at the southernmost end of peninsular Halifax at the entrance to Halifax Harbour (Fig. 1). The streets Young Avenue and Tower Road end in parking lots at the park; walk to the stops from these (Fig. 2). The park covers approximately 75 acres and the total distance around the stops is about 1.5 km. It takes about 2 hours to complete the trip. Stops 3 and 4 are 35 m above mean sea level, so there is quite a steep grade to them from the shore. There is no sampling allowed in the park.

Introduction

Point Pleasant Park contains some very interesting and spectacular sculptured erosional features resulting from geological processes during the Pleistocene era, the Ice Age. Erosional features can result from glacial ice scraping the bedrock as well as discharging meltwater sculpting the bedrock beneath the ice. On this field trip we will be making observations at four locations to see both of these effects on the rocks of Point Pleasant Park. The locations can be visited in any order and the descriptions reflect this.

Definitions and classification of erosional features

Ice contact erosional features:
The following schematic diagram (Fig. 3) shows several erosional features resulting from ice contact:

- polish—caused by intense scratching of bedrock by rock flour to a smooth surface
- striations (scratches, grooves, and gouges)—caused by rock fragments in base of ice abrading (gouging and scratching) the bedrock
- chattermarks and crescentic fractures—series of small curved fractures (chattermarks) or larger single crescent-shaped (hyperbolic) mark, concave to direction of ice movement and is made by a rock fragment in the base of the ice vibrating while moving (the vibration is similar to that made by the heal of a hand while being pressed forward and down on a flat surface) or single point pressure; there is no removal of any rock
- rock ledge (roche moutonnee, whale's back, and step)—asymmetric bedrock hill with a smoothly rounded or abraded upstream (stoss) side and a plucked irregular downstream (lee) side.

Later erosional features:
The following series of diagrams (Fig. 4) show several common water erosional features caused by coherent flow structures or turbulence. The pothole is the most common and familiar to many and is found in the bed of streams where water flows over bedrock. The other forms can be seen in the park at stop 2.

- pothole—near-circular, deep depression that may show spiralling, rising flow elements on their walls
- transverse trough—straight trough transverse to flow with a steep planar upflow face and a gentle somewhat eroded downflow face
- flutes, channels, furrows—depressions longer than they are wide, parallel to the flow direction with sharp to quickly rounded rims on the upflow direction, and in the case of the flute, pointing upstream; may be asymmetric and sinuous
- mussel-shell-shaped depression—a parabolic depression convex to flow direction with a sharp upflow and indistinct merging downflow rim
- sickle- or comma-shaped depression—a double or single armed depression with a sharp rim convex upflow and a crescentic main furrow downflow

Descriptions

Stop 1: Black Rock Beach
The rocky knob at the southeastern end of the beach is a rock ledge; the smooth northwestern slope is gradual and striated with a few glacial grooves, and the steep southeastern slope is sharp, fractured and uneven. The former reflects the scraping action of the basal part of the glacial ice and its entrained rock fragments on the bedrock. The latter face resulted from the plucking action of pieces of bedrock by ice aided by meltwater invading fractures and fissures, refreezing and expanding, breaking the bedrock.

The exotic pattern on the stoss slope is a low angle interference pattern between the ripple-marks and cross-beds in the sandy layers and the eroded surface. These ripples and cross-beds can be seen quite clearly when looking back into the outcrop from the lee side; the bedding dips approximately 30° to the north, almost parallel to the slope. These beds are on the south limb of a large east-west trending syncline that runs approximately through the middle of the container port. A poor, steeply dipping east-west cleavage associated with this folding event can also be seen in the outcrop. In addition there are several steeply dipping fracture surfaces oriented north-northwest that have facilitated groove development.

The rock has a pock-marked appearance; the pock marks cutting the cleavage surfaces in places indicating they are younger than the cleavage formation. These holes are the weathered remnants of the mineral cordierite, a metamorphic mineral that grew within the metamorphic aureole of the South Mountain Batholith which outcrops approximately 1 km to the west.

Stop 2: Chain Rock Battery

The rocky knobs here are also rock ledges similar to those at Black Rock Beach as well as rounded glacial mounds or roche moutonnée (here, more like whale backs than sheep backs). Striations and grooves oriented approximately 300°(NW-SE) are evident on the stoss sides, some very large, and the lee sides show evidence of plucking. Near the park benches beside the walkway is poorly developed glacial polish, seen as a dark spotting on the bedrock on the north sides of small bumps. The large trough shape of the Northwest Arm itself gives some evidence of the power of glacial erosion; the bedrock of the Arm is fractured, which facilitated erosion and straightening by the ice prior to the invasion by modern ocean waters.

Down by the shore near the park boundary is a poorly developed transverse trough and longitudinal channel (the axis is about 40° to the flow direction so is a combination of the two forms) with a metre-long comma-shaped depression and a mussel-shell-shaped depression (or sickle-shaped) on the lee side of the northernmost rock ledge. Current direction from the depressions indicate the water was flowing from northwest to southeast. Boyd et al. (1988) determined from the size of seven submarine tunnel valleys off Sable Island that at least 4.5 X 10^6 m^3/sec of water was running off from the melting glaciers on the Nova Scotia mainland during the main Laurentide ice retreat (present day Amazon River is discharging about 1.1 to 1.4 X 10^5 m^3/sec). This volume of water must have been eroding rock somewhere, evidence of what we may or may not seeing here.

The bedrock 'slate' is no longer slate-looking, in fact, it looks baked and there are more pock marks (weathered cordierite) than seen at Black Rock Beach. In addition to this, some fine-grained beds (formerly slate) show the mineral andalusite. Bedding is oriented from south to southeast dipping (260°/65° to 230°/65°) and is from distinct to partially disrupted or irregular (seen as irregular circular and elliptical zones of laminated meta-sandstones within the massive 'slate'). The poorly seen cleavage is nearly vertical, and oriented approximately NE-SW (235°/75°). Some of the circular areas that are rusty coloured are not bedding but are former concretions which result in a mineralogy different from the rest of the bedrock. All this is evidence that a large heat source is nearby and just as we suspect the South Mountain Batholith is about 400 metres away, on the other side of the Northwest Arm.

Stop 3: Price of Wales Martello Tower

The bedrock in front of the tower is a good example of glacial pavement, a smooth flat outcrop, ground by the base of glacial ice. At the southern end of the outcrop is glacial polish, seen as small (<5 cm^2) dark patches on the meta-sandstone beds which stand out in relief with respect to the slate. There are two directions of glacial striations here; one which is dominant and trending (NW-SE) 310°-320°, and the other of which there are a few examples trending nearly E-W (280°-295°).

You will notice in many areas that the pavement surface is offset several millimetres in height across fractures. This differential uplift may be due to uneven rising of the depressed bedrock surface during isostatic rebound after the glaciers retreated from the area.

The bedrock meta-sandstone layers are folded at this location forming S-shaped folds. This folding occurred during the mountain-building episode of the Acadian Orogeny, 400-365 million years ago.

Stop 4: Quarry Pond

The pond is a quarry that later filled in with water but the rock surfaces around the edges are natural. On the southeast side of the pond are large zones containing small (1-10 cm^2) black or dark coloured areas. These are glacial polish—formed by clay-sized particles and ice giving a smooth finish to the bedrock. Run your fingers along them to get the feel of their exceptionally smooth finish; there is no other surface like this in the immediate area. Some of the polished surfaces are ribbed. The polish is on the stoss side of small bumps (<0.5 cm high) indicating the glacier moved from north to south.

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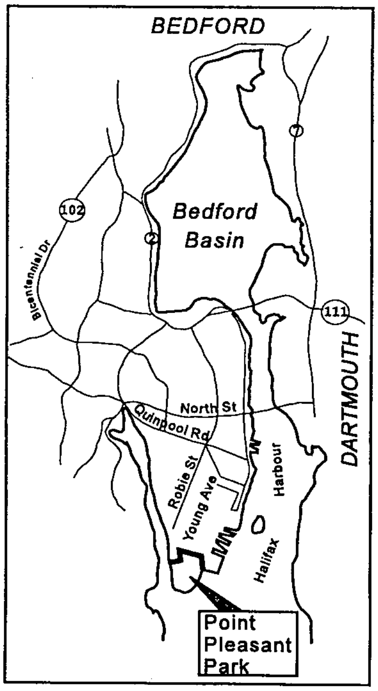


Figure 1: Location map for Point Pleasant Park, Halifax

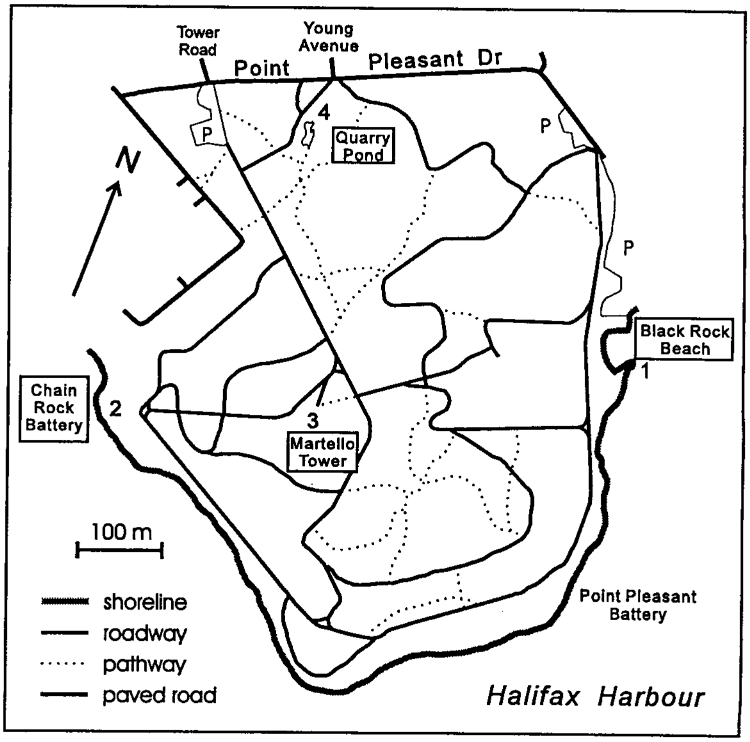


Figure 2: Location map for stops within Point Pleasant Park.

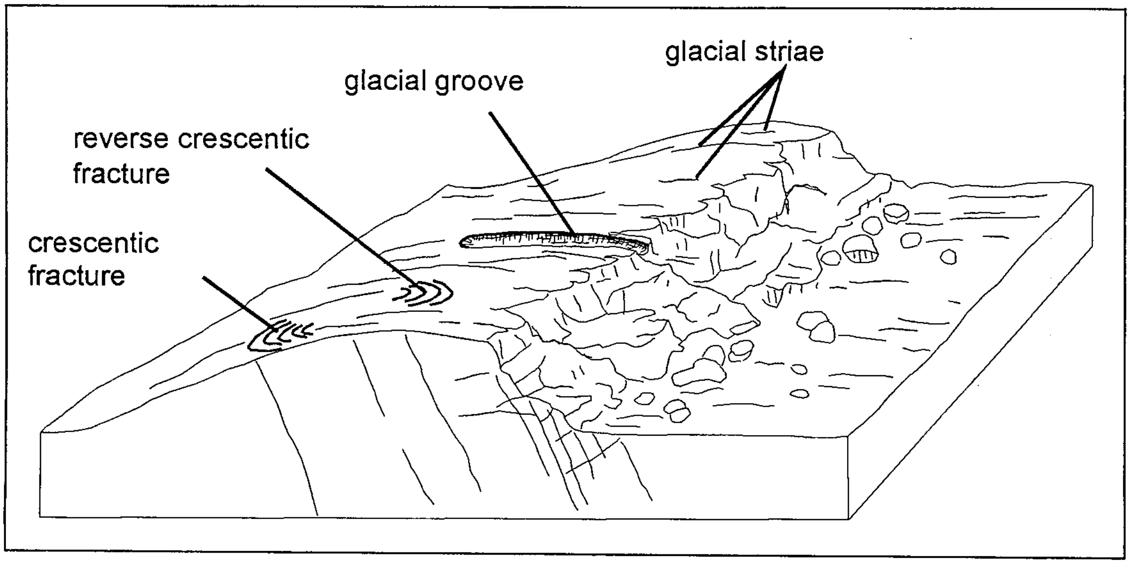


Figure 3: Generalized sketch showing ice erosional features.

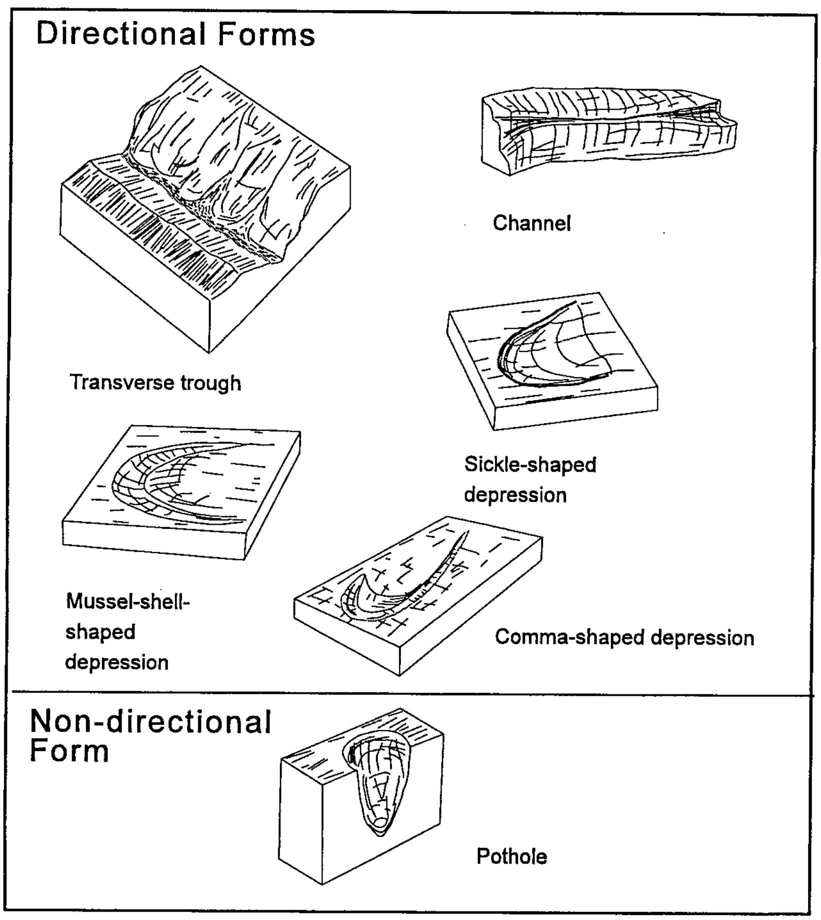


Figure 4: Generalized sketches of water erosional features.

Popes Harbour

The Weekend Dikes

Highlights

- Mafic (lamprophyre) dikes

Precautions

The Popes Harbour Weekend dike is on a series of exposed coastal outcrops which by themselves on good days are not too dangerous, but the exposure can be very dangerous in wet windy weather. Take extra clothing in preparation for quick changes in wind, and other conditions and wear strong sturdy boots. They are on low outcrops so are best viewed at low tide.

Directions

There are 10 Weekend dikes from Halifax Harbour to Sheet Harbour of which the East Jeddore, Little Harbour, Popes Harbour, and Sober Island locations can be driven or walked to. (Fig. 1). The others can be reached by boat. The Popes Harbour Weekend Dike (number 8 on Fig. 1) is the best exposed and one of the more complex of the 10, so it is the only one described.

All locations are on the Eastern Shore of Nova Scotia, off Route 7. The Popes Harbour dike is off the Coopers Road, in Tangiers (Fig. 2) and is by far the best exposed of the dikes. At the end of the Coopers Road, near to the fish plant, take the small road to the left and then near the end of this road is a walking trail (0.5 km) to the beach on Popes Harbour. The three northernmost outcrops of the dike can be reached from this beach, the others can be reached by boat.

Introduction

The Weekend dikes constitute a small "swarm" of 10 north-northwesterly oriented subvertical mafic (lamprophyre) dikes exposed along the Eastern Shore of Nova Scotia. All dikes have similar appearances and they show mineralogical, textural, and intrusive features typical of lamprophyres, i.e., forceful mafic intrusives with mafic phenocrysts and little or no feldspar that evolved from magmas rich in carbon dioxide, sulphur, phosphorous, and water vapour. Their consistent orientations and macroscopic features suggest that they represent a related lithologic subset of the Late Devonian mafic intrusions in southern Nova Scotia and individual dikes fall into two size categories.

Descriptions

The narrowest dikes at Borgles Island, Little Harbour coast, and Devils Island have ginger-brown weathering characteristics, they range in width from 0.01-1 m, and they display sharp and discordant contacts approximately perpendicular to regional east-northeast trending folds in the Goldenville Formation. Narrow dikes also have fine-grained prophyritic textures that result from chilling, and they finger out or show intrusive kinks and offsets along strike. At Borgles Island, three narrow dikes occupy late-Acadian, northwesterly-oriented shear fractures that presumably represent the magma conduits. The Devils Island dike shows strong propylitic alteration (replacement of horneblende and biotite to chlorite, calcite, sphene, and iron ore by hot migrating waters containing CO_2) similar to dikes seen at the Ovens.

Wide dikes at Sober Island, Popes Harbour, Pleasant Harbour, Little Harbour Road, and East Jeddore range in width from 12 to 15 m, they have melanocratic (>60 per cent mafic minerals) appearances and either resist weathering (at Sober Island and Pleasant Harbour), or weather preferentially (at Popes Harbour). All examples show medium-coarse grained, porphyritic textures and have fine-grained chilled margins that form approximately 10 per cent of the total dike width. Contacts with the Meguma Group appear straight over tens of metres, and contact metamorphism within the country rocks occurs as insignificant bleaching and the annealing of tectonic cleavages immediately adjacent to the dike margins. The Sober Island and Pleasant Harbour dikes occur proximally to *en echelon* 20-70 cm companion dikelets that presumably represent small, upwardly penetrating fingers of dike material that connect to the main dike at depth. The Pleasant Harbour and Popes Harbour dikes contain irregular pods and stringers of fractionated, feldspathic lamprophyre, and late-stage mobilisation of this material caused autobrecciation at the centre of the Pleasant Harbour dike. The anomalous Popes Harbour dike represents a composite body, comprising two outer chill zones approximately 1-2 m wide and separated from a 10-12 m inner zone of xenolithic lamprophyre by a prominent contact. Up to 30 per cent exotic metabasite and metapelitic xenoliths and xenocrysts of lower crustal derivation occur in the dikes at Popes Harbour and Pleasant Harbour, and the chilled margins contain ovoid, 1 cm diameter quartz-orthoclase-calcite-epidote-pyrite felsic globular structures.

Petrography

The phenocrysts in the wide, coarse-grained dikes consist of elongated brown amphibole (29-51 per cent) and diopside (0.5-13 per cent). They show crude parallel alignment and are euhedral to subhedral shaped, that is, bound completely by crystal faces to partly bound by crystal faces and partly bound by irregular surfaces. Rare, ovoid chlorite mats enclosing chrome spinal within the Popes Harbour, Sober Island, and Little Harbour road dikes may indicate the previous presence of olivine. Plagioclase laths range in size from 0.2 to 1 mm, and they occur only in the groundmass where they nucleated with equant amphibole, clinopyroxene, quartz (0.7-9 per cent) and K-feldspar (2-15 per cent). Biotite (~17 per cent) occurs in the central zone of the Popes Harbour dike, where it represents a phenocryst phase that apparently precipitated partly at the expense of amphibole (~9 per cent). Patches of anhedral calcite, epidote, and chlorite (~5 per cent), coalesce to form pseudomorphs after plagioclase and amphibole phenocrysts at the centres of large Weekend dikes, and the accessory minerals include rutile, perovskite, and sphene.

Preliminary Interpretations

Euhedral to subhedral textures and evidence of reactions between primary minerals and water solutions resulting in calcite, epidote, and chlorite are diagnostic of lamprophyres. The occurrence of augite, diopside, and hornblende as the mafic phenocryst phases, and plagioclase as the dominant groundmass feldspar, confirm the spessartite lamprophyre classification for the Weekend dikes. Also, the presence of biotite in the Popes Harbour dike imparts kersantitic or "dioritic-looking" characterstics similar to those of the Forbes Point and Mcleods Cove lamprophyres. The absence of contact metamorphism, the development of felsic globular structures, and the elongate amphibole habits with weak zoning, all suggest rapid emplacement and cooling. This rapid undercooling in the narrow dikes probably preserves the early anhydrous crystallizing assemblage. The importance of amphibole over clinopyroxene in all the larger (presumably slower cooled) dikes, suggests that increased volatile exsolution occurred in the lamprophyric magma, probably resulting from decompression during dike emplacement. Silica (atomic) concentrations between 6.9 and 7.2 in amphibole suggest that they crystallised from intratelluric mafic magmas and wide compositonal zoning in plagioclases also suggest a predominant mafic magma containing volatiles rather than a purely volatile medium considered plausible for lamprophyres. The smallest Weekend dikes intrude Acadian crustal extensional weaknesses ("shear joints"), and the larger dikes and other Lower Devonian Mafic Intrusions may also mimic this intrusive behaviour.

Source

Tate, M.C., 1995. The Relationship Between Late Devonian Mafic Intrusions and Peralumnous Granitoid Generation in the Meguma Lithotectonic Zone, Nova Scotia, Canada. PhD Thesis, Department of Earth Sciences, Dalhousie University, Halifax, Nova Scotia, 528 p.

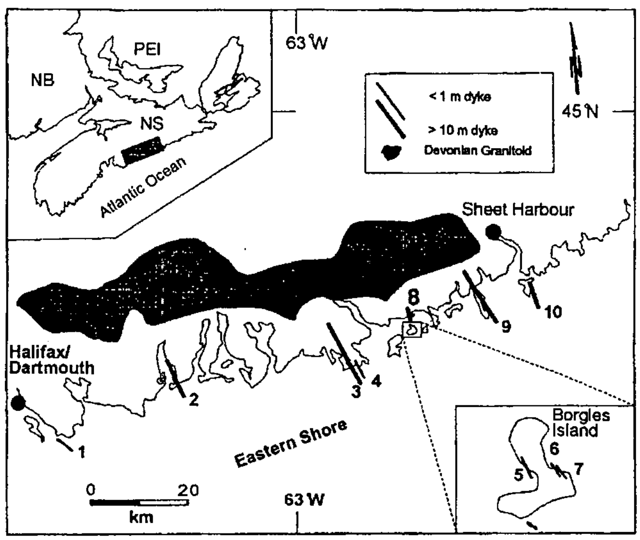


Figure 1: Map showing locations of the Late Devonian mafic dykes along the Eastern Shore. The Popes Harbour Dyke is number 8. From Tate, 1995.

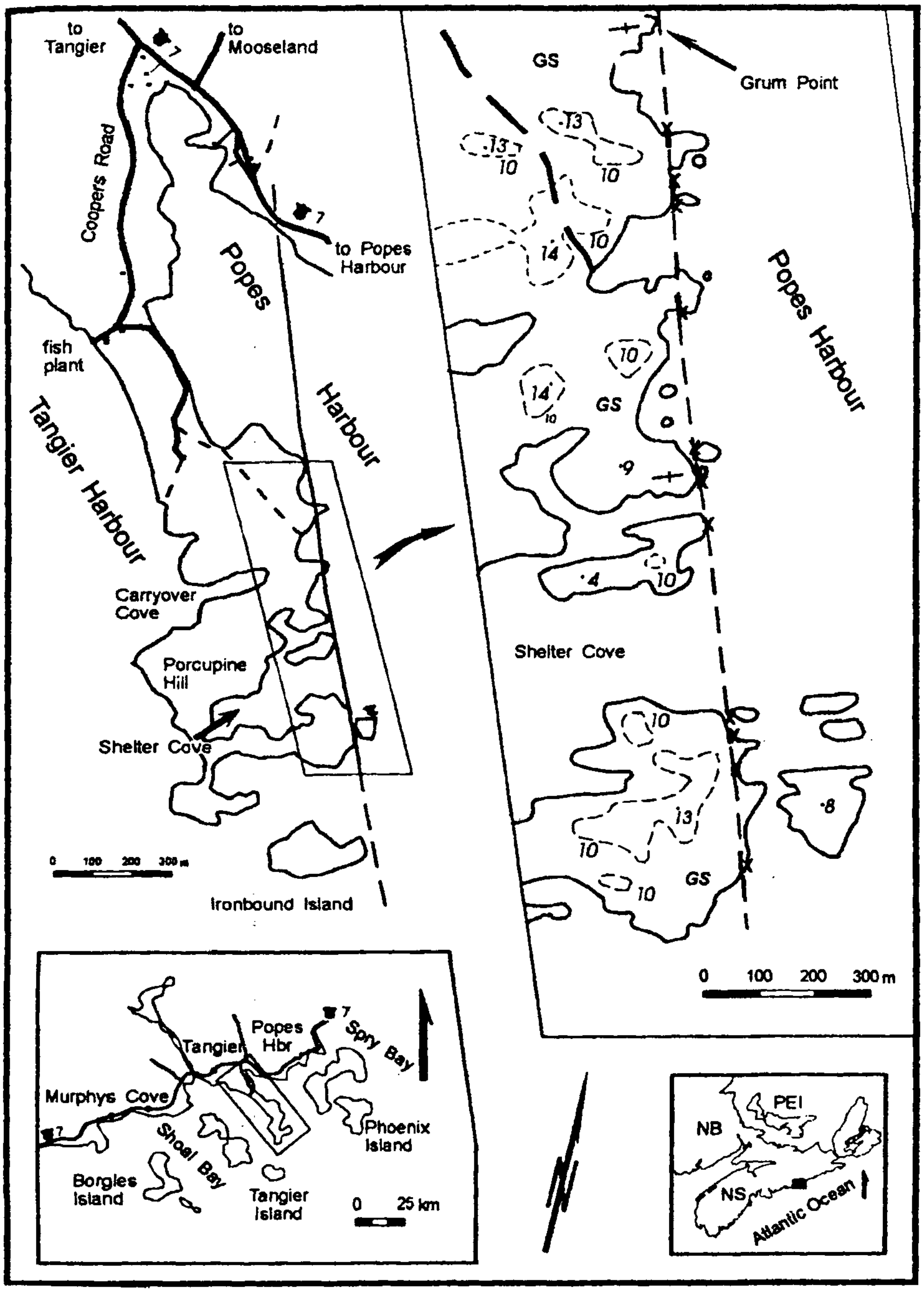


Figure 2: Detailed location map for the Popes Harbour mafic dyke. contours are in metres. From Tate, 1995.

Wine Harbour

Deformation and Quartz Veining in the Meguma Group

Highlights

- slaty and spaced solution cleavage in meta-pelites and sandstone
- history of complex quartz veining and deformation
- relationship between cleavage and regional folds in the Meguma

Location

Starting at Sherbrooke, travel south towards Sonora, taking the paved road past Sherbrooke Village and mill. Past Sonora, the pavement changes to gravel. Continue, taking the first right fork towards Wine Harbour.

Alternately, if you are going south on Route 7, it is easier and quicker to take the Route 211 turnoff and go past Indian Harbour Lake. Then turn west onto the gravel road to Port Hilford, take the next 2 left turns, and end up going towards Wine Harbour. Either route will take you to Wine Harbour settlement.

At Wine Harbour Settlement take the left fork towards the Barrachois (rocky beach). Park at the edge of the wharf, the rocks we are examining stretch along the shore northwards for about 500 m. The woods to the west contain the remains of old gold workings: be careful—the pits are not covered up.

Introduction

Mainland Nova Scotia is divided into two separate tectonic zones: the Avalon domain to the north and the Meguma domain to the south. These are separated by a down-dropped block and called the St. Marys–Guysborough Fault. The geology of the Meguma domain is distinctly different from the Avalon and this difference allows us to separate them into two tectonic domains.

Although the details of how the Avalon and the Meguma domains came together is highly debatable, it is known that they must have originally formed far apart in different geological environments. We also know that they must have been brought together before the middle of the Early Carboniferous because the rocks of the Windsor Group and the Upper Horton Group extend across the domains.

The rocks of the Meguma domain comprise metasediments of the Cambro-Ordivician aged Meguma Group and igneous plutonic rocks of Devonian age (approximately 370 Ma). The Meguma Group consists of metamorphosed sandstones and mudstones (quartzite, meta-sandstones, meta-siltstones, and slates) of the Goldenville Formation and black slates of the Halifax Formation. Although the Halifax Formation is structurally younger than the Goldenville Formation, in any specific area of the Meguma, exact stratigraphic relationships are difficult to determine. Both formations appear to represent deposition from marine turbidites.

Along the Eastern Shore, the Meguma Group has undergone a complex deformational history. These deformations can be seen as discretely different cleavages and the number and intensity of deformations increase northward from the Atlantic Coast to the St. Marys-Guysborough Fault. At the coast we will be examining two types of cleavages, a slaty cleavage in the slates and a spaced cleavage in the coarser units.

Description

The exposures at Wine Harbour (village) overlook Indian Harbour to the east. Here a stratigraphic thickness of approximately 45 m of interbedded metasandstones, meta-siltstones and slate are intruded by a series of gold-bearing and barren quartz veins that are part of the Wine Harbour Gold District. Between 1862 and 1939, 42,726 oz of gold was extracted from approximately 80,000 tons of rock, much of it from the Plough Lead Belt west of the outcrop. Most old workings are now inaccessible but gold in quartz can still be found around the old dumps and shafts above the outcrop. The beds here dip 80° south and are approximately 300 m from the fold axis, which lies to the north. There are two distinct morphological varieties of cleavage present, one a slaty cleavage in the slates, and the other a spaced (pressure solution) cleavage in the metasandstone. The presssure solution cleavage exhibits a convergent fan with respect to the Wine Harbour anticline and the strike is oblique to the fold axial trace. The slaty cleavage forms a divergent fan in the same succession and its trend is subparallel to the fold axial trace. In some areas the cleavage slowly refracts from bed-to-bed indicating grading. In other areas the change in strike is abrupt.

Quartz vein arrays of pre- and post-major folding age are particularly well displayed at Wine Harbour. These comprise stratiform, stratabound, side, step, discordant and vertical veins. From the the descriptions below, see if your can identify each type of vein on the outcrops. While doing this decide which veins are pre-, syn- and post-deformation by looking at their relationship with the cleavage and from crosscutting relationships of individual arrays. The attitudes and angles between bedding, the cleavages, the bedding-cleavages lineations, the veins, the fold axes and axial planes do not form a simple geometric relationship. At this outcrop, the discordant and vertical veins contain less than 10 ppm Au while the others contain up to 600 ppm Au.

Vein type	*Description*
Stratiform	alternating layers of qtz ± carbonate ± clays often with selvages of host rock; parallel to bedding, in slates, auriferous and with sulphides
Stratabound	pinch-and-swell, inclusions of green chloritized slate; locally cross-cuts bedding; cuts all rock types but restricted to single bed horizons; some with bilateral wall rock alternations consisting of arsenopyrite + sericite ± tourmaline with large arsen-opyrite pseudo-porphyroblasts with fibrous quartz in pressure shadows
Step	discordant in a series of folded steps traversing bedding, massive quartz with little deformatin in coarse units and pinch-and-swell with muscovite K-feldspar sulphides ± carbonates in slates
Side	tightly folded tapering away from T-junctions withother veins, widely variable.
Discordant	crosscutting folded tension gashes restricted to coarse units
Vertical	vertically oriented with qtz chlorite ± K- feldspar ± muscovite ± sulphides, northerly trending and can occur as enechelon sets.

Go up the shore to the north and at the next major outcrop, find bedding and cleavage. The vertical cleavage is the dominant structure and should not be confused as bedding; the bedding is difficult to see because it is near horizontal to shallow south dipping. There are a few carbonate-rich zones (concretions) and indistinct bedding laminations on some outcrops to aid you. Continue up the shore to the next big outcrop where bedding becomes more distinct again. In this part of the walk the major Wine Harbour fold axis has been traversed, so the bedding should have appreciably changed its attitude to north dipping. Again note the two different cleavages and their relationships. The pressure solution cleavage is convergent and the slaty cleavage divergent.

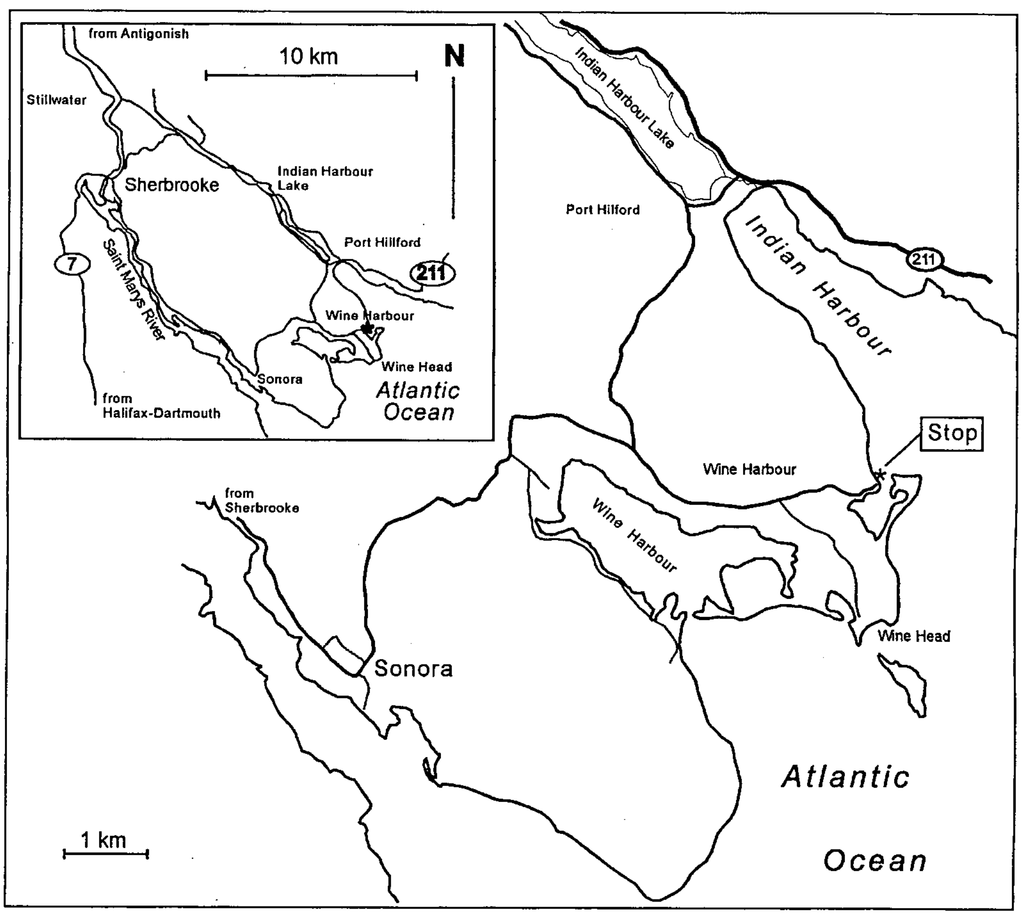
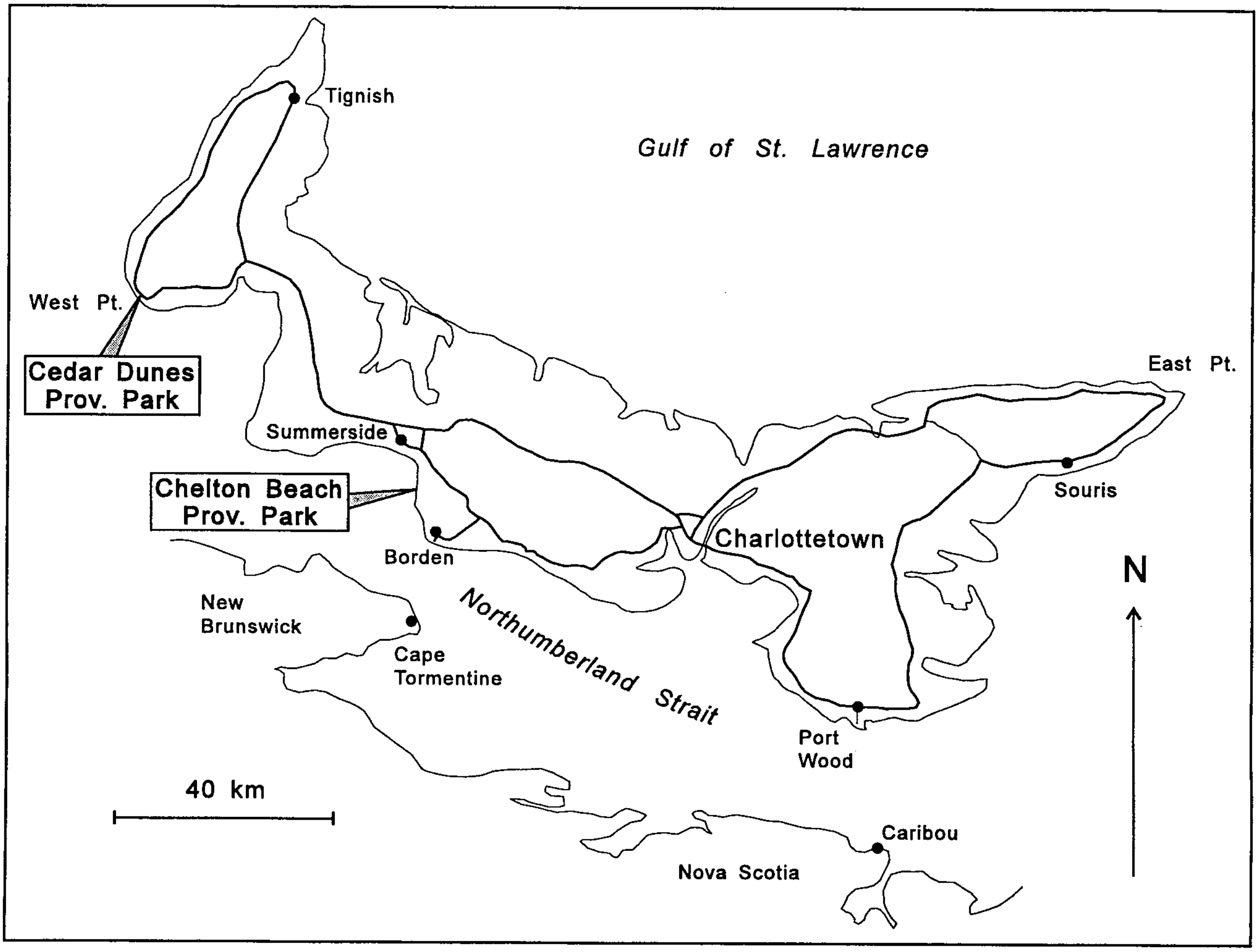


Figure 1: Location map for Wine Harbour.

PRINCE EDWARD ISLAND



Outline map of Prince Edward Island showing locations of field trip descriptions.

Chelton Beach Provincial Park

The Red Beds of Prince Edward Island

Purpose

The bedrock at Chelton Beach contains excellent exposures of the coarse-grained basal- to middle-beds of the Kildare Capes Formation of the Pictou Group redbeds.

Location

There are several ways to get to Chelton Beach Provincial Park (Fig. 1). If you are coming from Borden turn north on Route 10. After 5.5 km, at the intersection, stay on Route 10 by veering left and at 8.3 km take the paved Chelton Shore Road going west. This is the first paved road 2.8 km past the intersection. If you are on Route 2 turn south on Route 10 at Central Bedeque and then 1.5 km past the sharp curve in Bedeque take the Chelton Shore Road going west. Keep on the Chelton Shore Road making the sharp curve at 2.3 km. The entrance to the park is approximately 1.2 km past this curve. Walk down to the beach, and go left to view the outcrops.

Introduction

During Late Carboniferous to Permian time much of the Maritimes including all of Prince Edward Island was part of the Carboniferous Basin, the site of fluvial deposition by material eroded from the Appalachian Mountains to the west. Grey sandstone and conglomerate interlayered with red to green shale and minor coal were deposited in New Brunswick whereas later and to the east, redbeds were deposited in the region of PEI.

On Prince Edward Island the sediments are subdivided informally (van de Poll and Ryan, 1985) into two major facies sequences as shown in Fig. 2. In the west and central regions is the New Brunswick Platform where normal or average rate of sediment accumulation has taken place and the sediments are submature (high quartz/quartzite content, sub-arkosic wackes, and well-rounded clasts). To the east is the Cumberland sub-basin, an area that underwent an accelerated period of subsidence and sediment accumulation which resulted in less mature, highly arkosic wackes with less well rounded conglomerate clasts than those of the Platform.

In addition to this (see Fig. 2), the rocks of PEI are subdivided on composition, maturity, and grain size variations into four fining-upwards megasequences, several hundreds of meters thick. They have a coarse-grained conglomerate base, medium- to fine-grained middle and a fine-grained to mudstone top. These four megasequences were promoted to formational status by van de Poll (1989) and are considered part of the Pictou Group.

The sediments were transported and deposited by river systems over an extensive alluvial plane sloping gently to the east with more humid conditions giving way to arid. Drilling for oil, coal, and gas has produced rocks which have yielded spores ranging in age from Stephanian (late Pennsylvanian) in the west to Sakmarian (early Permian) in the east and northeast. The rocks of the Chelton Beach area (Kildare Capes Formation) are centred on the approximate boundary between the Carboniferous and Permian Periods. Slight basement movement as well as salt/gypsum diapirism of the Windsor Group has caused some of these beds to be folded and/or dipped gently to the northeast but overall they are gently inclined to flat-lying.

Chelton Beach

The rocks of Chelton Beach area belong to the Kildare Capes Formation (van de Poll, 1989). The base of the formation is to the west by Seacow Head and is marked by conglomerates with a high rhyolite clast content overlying distinctly finer grained beds of the Egmont Bay Formation. The strata dips gently (approx. 3^{0}) to the northeast and fines upwards to the coarse basal beds of the Hillsborough Formation (in the east near the Hillsborough Bay region), the third megasequence. The megasequence is estimated to be about 350 m thick.

At Chelton Beach the rocks are deep red to brown coarse- to fine-grained sandstone with minor amounts of sandy conglomerate and siltstone. The beds have irregular bedding planes giving shapes from weakly tabular to concave lenses averaging about 20 cm thick (up to 1 m). Some beds are channelled into the lower ones. Internally the beds contain laminations and low-angle cross laminations varying in thickness from a few mm to 3 cm in the coarse beds and <4 mm in the finer beds. In the some of the coarser beds the laminations are pebbly sandstones and appear as reduced zones (non-oxidation of the hematitic matrix) greenish in colour. Clasts rarely exceed 20–25 mm in diameter, most are 2–6 mm, are well-rounded, while compositions are dominated by quartz/quartzite, granite, and rhyolite with most sand-sized grains dominated by quartz, feldspar, and muscovite. There are very rare current lineations and ripple marks. The cement is carbonate material. The rocks are interpreted to represent levee, overbank or crevasse splay, and minor scour-channel deposits on a silty floodplain.

Some of the sandstone beds are chaotic or disrupted and are interpreted to have been altered by dewatering processes after deposition. This is probably a response to high pore-fluid pressures due to either sediment loading, buoyancy or compaction.

The bedrock is cut by vertical joints spaced 1– 3 m apart as well as longer ones that are irregularly spaced and randomly oriented. Some joint surfaces are coated with Mn oxides. The outcrops form cliffs 2 –4 m high and are capped by 1–2 m of thick till.

References

van de Poll, H.W. 1989. Lithostratigraphy of the Prince Edward Island redbeds. Atlantic Geology, vol. 25, no 1, pp. 23-36.

van de Poll, H.W. and Ryan, R.J. 1985. Lithostratigraphic, physical diagenetic and economic aspects of the Pennslyanian-Permian transition sequence of Prince Edward Island and Nova Scotia. Guidebook, Geological Association of Canada–Mineralogical Association of Canada, "Fredericton '85", Field Excursion 14.

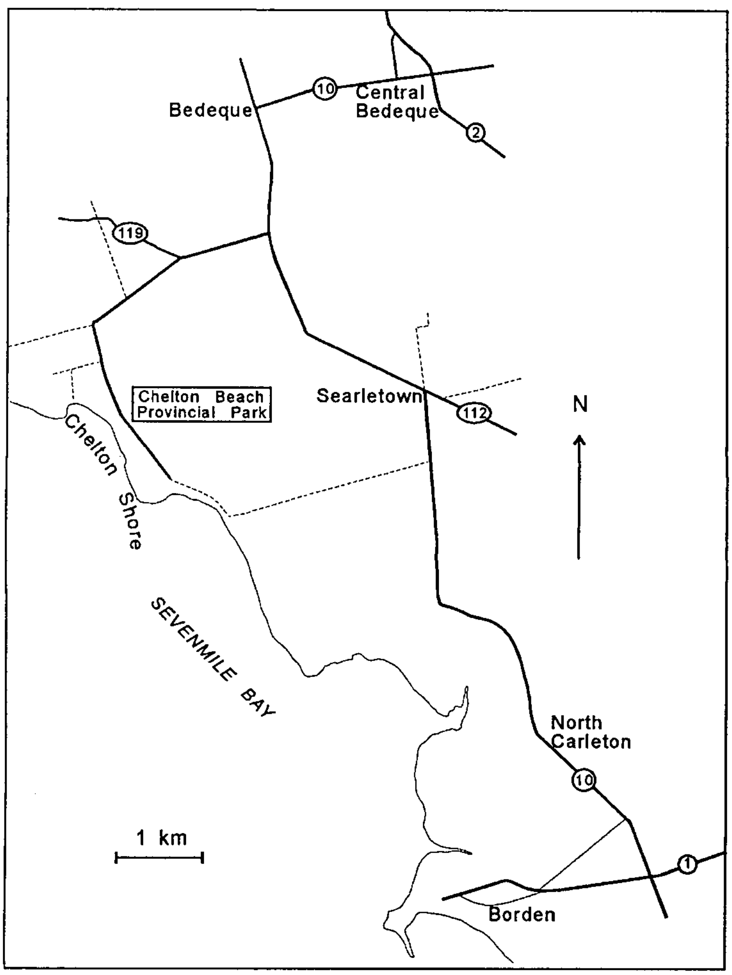


Figure 1: Location map for Chelton Beach Provincial Park, PEI.

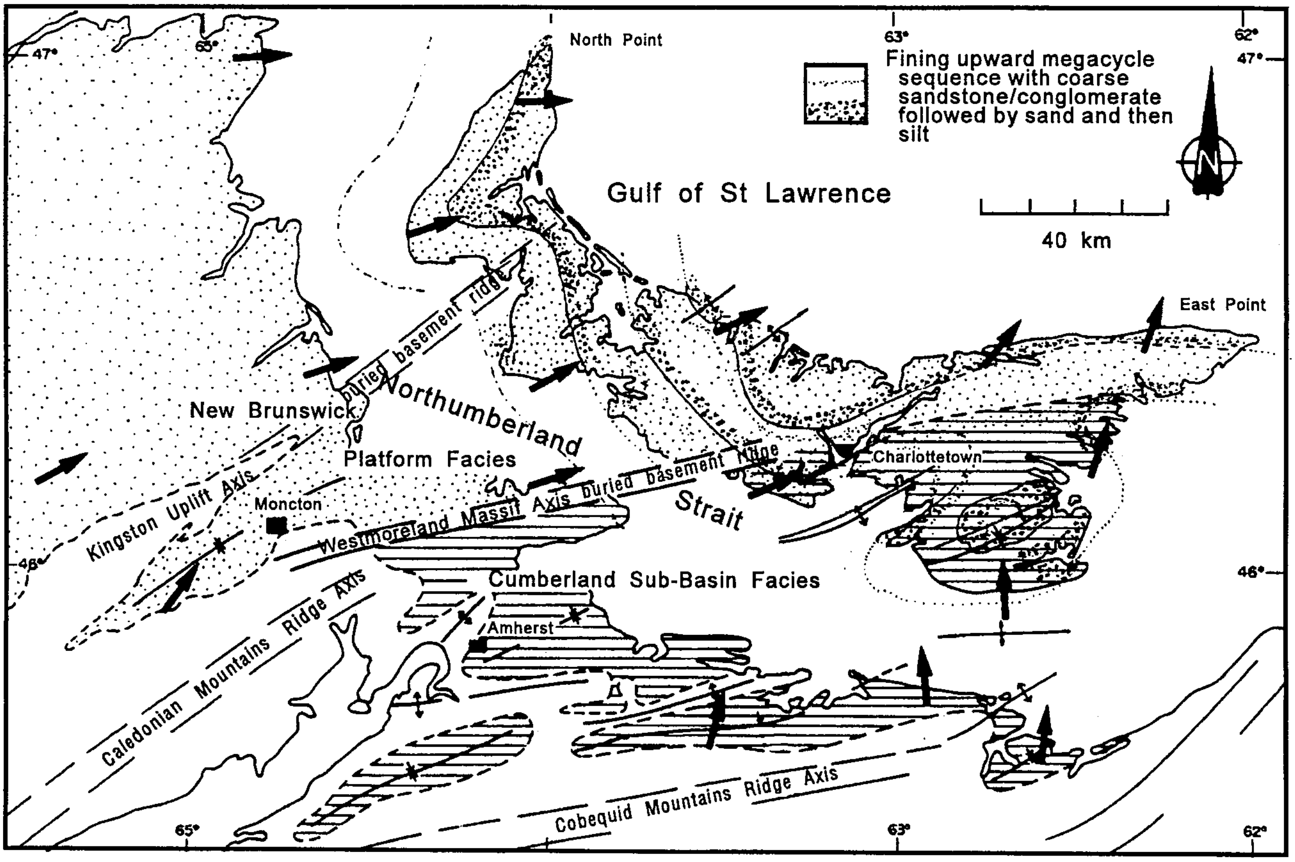


Figure 2: General geology map of Prince Edward Island and adjacent areas of New Brunswick andNova Scotia. Modified from van de Poll and Ryan, 1985.

Cedar Dunes Provincial Park

Purpose

The area shows how coastal processes affect geological and biological features on modern beaches and how they change over time.

Location

Cedar Dunes Provincial Park is located on the western end of Prince Edward Island, on Edgemont Bay (Fig. 1). Take highway 2 west to 14 and then go to West Point.

Introduction

About half a kilometre east of the West Point Lighthouse, Mr. B. B. Dunville built a lobster cannery in 1906, on piles, so boats could sail beneath it for unloading. By 1976, the foundation of the cannery was 143 metres from the high water mark on the beach. That is, the beach advanced at an average rate of 2 metres per year during that period. This is about half of the 4 yards (3.6 m) per year estimated by Mr. Dunville for the 26 years prior to 1948.

Beaches

A beach is the accumulation of loose material along a shoreline due to wave action and extends from the offshore where waves first begin to move material on the bottom to the extreme upper limit of wave action. The area below the low tide line which is never exposed is termed the offshore (Fig. 2); this region may contain sand bars, or shoals, which may be partially exposed at extreme low tide. The part of the beach exposed at low tide, but submerged at high tide is the fore shore, or intertidal zone. Above this zone is the backshore, which is only affected by waves during extreme conditions, during unusually high tides or storm surges. The backshore is generally composed of one or more berm s–horizontal or landward-sloping terraces.

The coastline shape, the sediment supply, and the type of energy all play a role in beach formation. In eastern Canada the coastlines are generally submergent, that is, sea level is rising relative to a fixed point on land. Many bays have been formed by flooding of river valleys and the sand bars across their mouths are the result of erosion of headlands and the subsequent deposition of much of this material in these areas. Sediment is generally in constant supply from the headland erosion and in order for the beach to persist, the amount of sediment being deposited must equal the amount being eroded or removed. In Prince Edward Island the glacial till as well as the weakly consolidated Late Carboniferous--Early Permian aged sedimentary rock provides material at a tremendous rate but very few beaches receive sufficient sediment to build seaward, or even to keep pace with rising sea level in the long term.

Energy is supplied to beaches in the form of waves, currents, and tides. In western Prince Edward Island storm generated wind waves supply by far the greatest amount, with tidal currents second. Wave size increases with wind speed, duration and area over which the wind blows (known as the fetch); the fetch from the northwest is over 200 km in this area. As waves approach the shore they slow down where they begin to first "touch" the offshore zone, steepen, and eventually break. Long-continuous waves usually approach a beach at an angle and as a result tend to bend or refract parallel to the shoreline. This causes a current down the shore, parallel to the beach in the same direction as the approaching waves, and is known as the longshore current. The longshore current can also transport sediment along the offshore. Longshore drift, the swashing of waves up the beach face at an angle, and then backwash perpendicular to the shoreline, transports sediment along the beach in the same direction but in a zigzag pattern.

The interaction of these processes usually produces primary coasts, those affected mostly by subaerial geological processes such as streams, glaciation, volcanism, and tectonics. In western Prince Edward Island however the coast from North Point to West Point is of secondary origin. Here the shoreline is shaped mainly by marine wave erosional processes, it has become smooth and straight with a balance between the energy of the waves and the configuration of the shoreline. The processes shaping the primary features (bays, estuaries, etc.) have been overtaken by the coastal offshore processes.

Dunes and Vegetation

Many beaches are backed by extensive dunes, regular shaped heaps of sand above the high tide mark. Plants tend to stabilize and bind the sand particles in a dune and prevent their migration or destruction. The vegetation changes in a predictable and progressive way over time according to different conditions and levels of resources (nutrients): primary change or succession occurs when plants invade a new unvegetated area and secondary succession occurs when vegetation is disturbed but the soil remains. Disturbance in this case usually means deforestation by natural causes but also through human activity. A general beach profile showing this succession is depicted in Fig. 2.

The earliest successional stage is represented by salt tolerant species that grow on the beach just above high tide such as the Sea-Rocket (*Cakile edentulata*), an annual that grows new from seed each spring, dying over the winter. The next successional stage is the plants growing on the dunes themselves binding the grains around their roots and stems, usually grasses and the beach pea (*Lathyrus japonicus*). Further from the shore soil becomes developed and diversity increases. Plants such as goldenrods (*Solidago* sp.) and asters (*Aster* sp.) grow along with isolated spruce trees and shrubs such as bayberry, wild rose, juniper, blackberry, blueberry, and cranberry. Eventually more and more white spruce trees become established and grow larger, whereas further inland red spruce is more abundant. Near bogs where the soil is wet black spruce and cedar grow. The trunks and branches of many of these trees may be covered by lichens and the soil covered by mosses.

Description

The sediment at West Point consists of sand and gravel dominated by quartz grains and rock fragments. Supply of these materials comes from erosion of the shoreline to the north between West Point and Carey Point where the beaches are actively eroding and migrating inland, but reports that the West Point lighthouse had been moved inland several times have not been confirmed by the Department of Transport. The material is transported to West Point by longshore currents and drift; according to local fishermen, for three miles offshore the longshore current is sufficiently strong to cause large quantities of gravel to drift across the bottom. When the current sweeps around West Point and across Edgemont Bay, an eddy, which carries sand away from the main longshore drift but is not able to keep the sand in suspension, drops the material to the east of the point. As a consequence the beach is building out to the south-southeast as a series of ridges.

The diagram (Fig. 2) illustrates this but you can also see this situation on the beach by studying the refraction of waves around the point. The longshore current sweeps past West Point and across Edgemont Bay and the right-angle bend creates the eddy. The dashed lines on the diagram represent former positions of the beach, recognizable on aerial photographs. The former beaches are a series of ridges accentuated by zonation of vegetation; on the tops of the ridges are mainly grass, in the troughs are shrubs and broad-leafed plants and berries. Some conifer trees are natural but many have been planted by parks personnel to try to stop erosion.

On the western shore some dunes have been eroded and are exposed behind the beach. Inside the dunes you can see fine laminations that are either horizontal or at an oblique angle (known as cross beds) indicating that the sediment was deposited on a slope. On the back sides of the dunes and in places on the beach ridges are small wind-generated asymmetric current ripples (one side longer and more gentle than the other side) and in the offshore and intertidal area are wave-generated symmetrical ripples as well as current ripples.

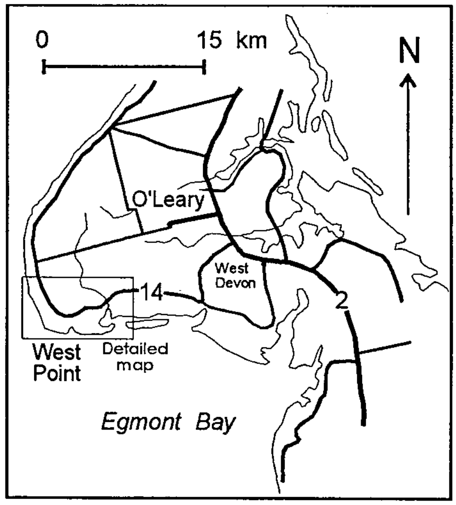
Behind the park and lighthouse is a dune-cedar area. In the low areas the land is swampy, moss covered, and has cedar trees. On the ridges where there is better drainage are broad leafed plants and ferns.

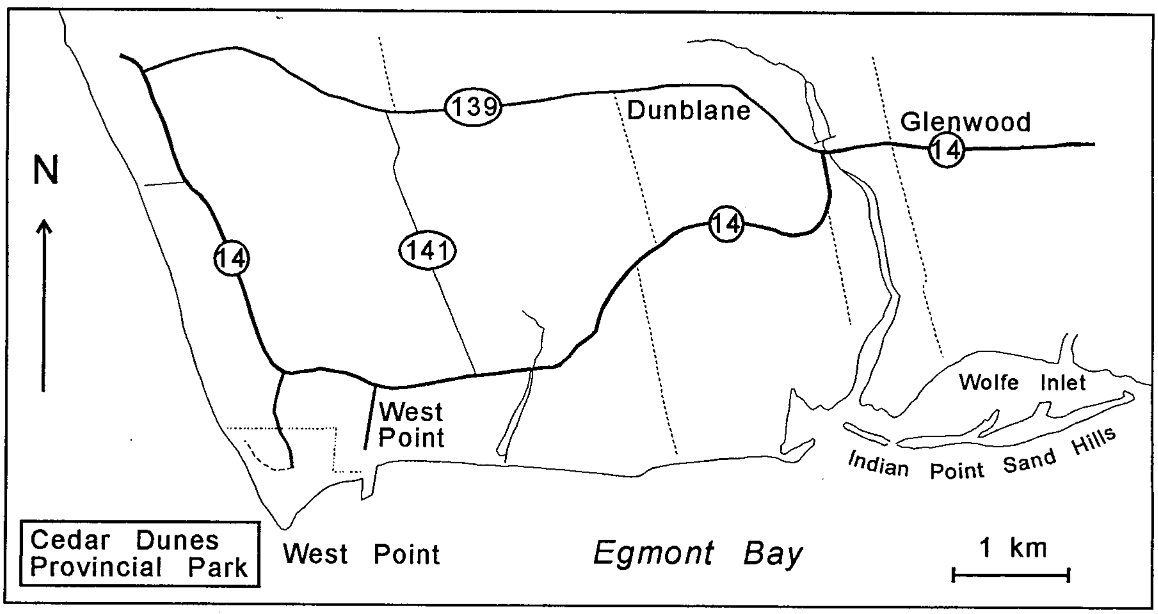
Exercise

Deposition is sufficiently rapid that this could be used as an exercise that should show measurable results during the time the students are in high school. Suitable permanent marks, well back from the beach, could be used to measure the position of the high water mark at suitable intervals, say every 6 months. In combination with a biology class, the development and migration of vegetation zones on the beach could be monitored; judicious placement of narrow trenches should show bedding, variations of sediment size, and other depositional features. Suitable record keeping should also produce, over a period of some years, a record that might make a suitable science project for a class, that is, to analyze the accumulated data and draw more detailed conclusions than those of the above brief outline.

Source

Adapted from an idea supplied by Dr. G.C. Milligan, Emeritus Professor, Earth Sciences Department, Dalhousie University, Halifax, Nova Scotia, B3H 3J5





Figures 1: Location maps for Cedar Dunes Provincial Park, PEI.

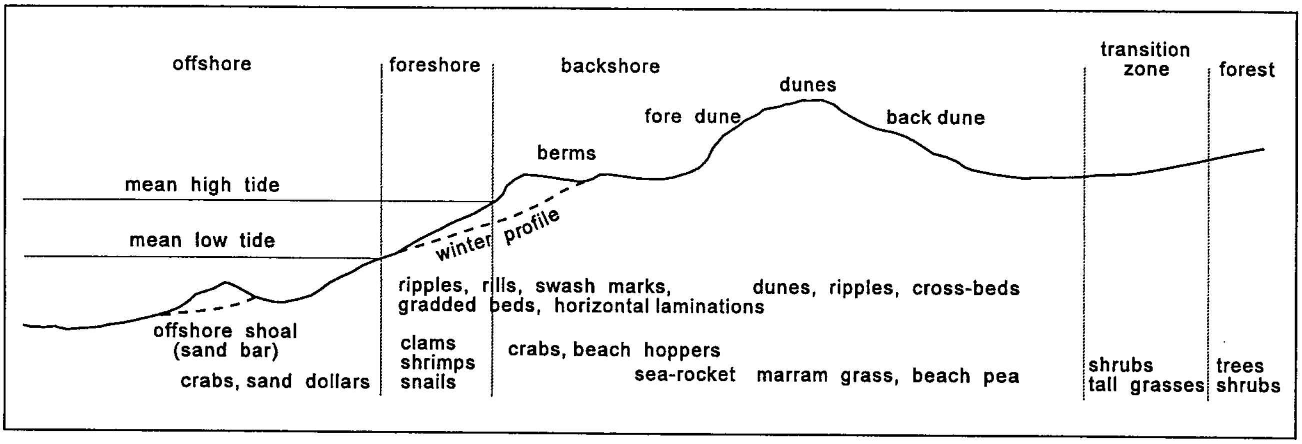


Figure 2: Generalized beach profile showing physical and biological characteristics

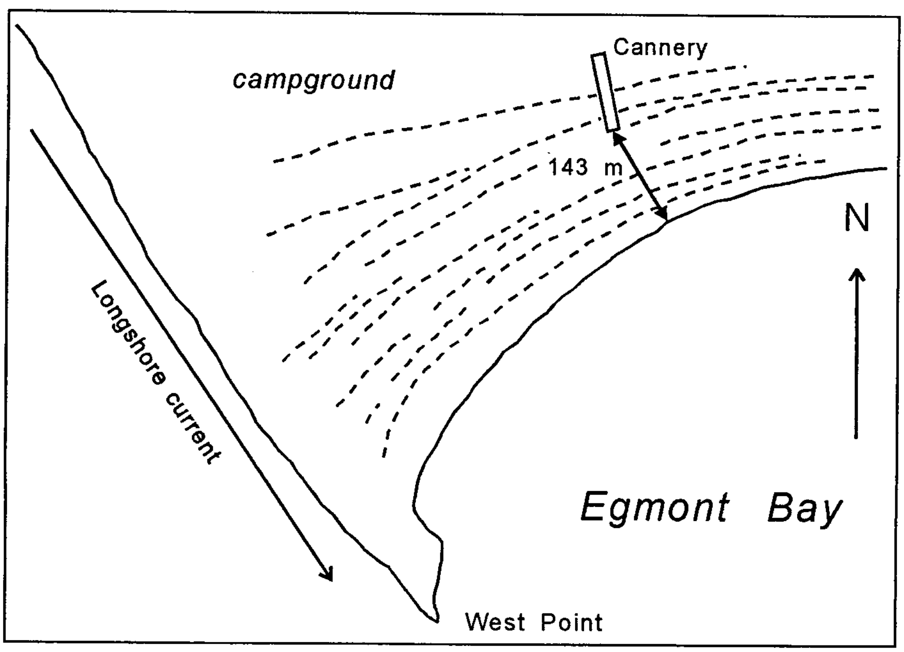
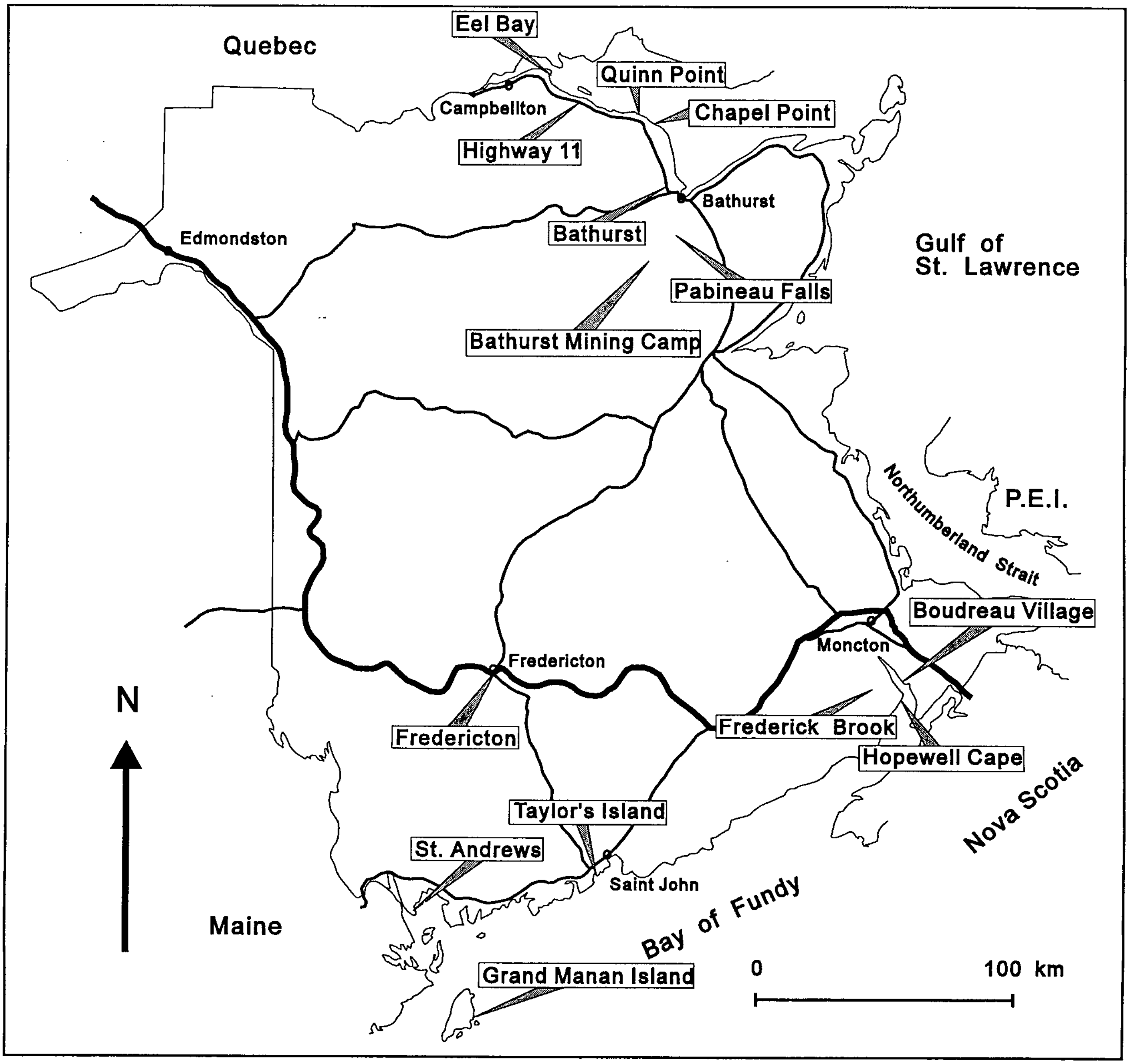


Figure 3: Detailed map showing beach ridges at West Point. Cedar Dunes Provincial Park. Drawn from air photos. After Milligan, 1980.

NEW BRUNSWICK



Outline map of New Brunswick showing locations of field trip descriptions.

Boudreau Village

Cyclic Fluvial Deposition of the Albert Formation

Highlights

- important aspects of the Albert Formation:
- fluvial channel facies
- interchannel/overbank (delta plane) facies
- alluvial fan facies
- upper contact with Weldon Formation

Directions

Proceed from Moncton towards Dieppe on Main Street. Turn southeast on Route 106 towards Fox Creek (marked as Route 6 or Acadia Avenue) which is 10 km away. Continue past Fox Creek for 1 km and then turn right onto Route 925 (Fig. 1) to Dover following signposts to Gautreau, Belliveau Village, and Boudreau Village. At Pre-D'En-Haut, 13.7 km from the 106/925 intersection, take the shore road towards Belliveau Village, leaving the 925 which goes east. At the intersection in Belliveau Village, 3.5 km from the 925, keep to the shore road and a further 1.2 km brings you to Boudreau Village. Park at the northern edge of Boudreau Village, across the small distributary channel marked "14 Aboiteau des Boudreau" (Fig. 1). Across the river is an old gypsum mine and mill. Take the trail to the beach and walk south along the Petitcodiac River to the outcrops.

Precautions

This stop is tidal dependent. Be sure you know the state of the tide at all times and do not wander out onto the wet sticky intertidal mud. Caution must be exercised at all times and never venture out into an incoming tide.

Description

The local geology of the Boudreau Village section is indicated in Fig. 2 and the generalized stratigraphic section of this location is indicated in Fig. 3. The base of the exposed sequence in the south part is composed of three small-scale fining-upward fluvial channel and associated overbank deposits, as illustrated in detail in Fig. 4. A variety of sedimentary structures may be observed, including straight to smooth crested ripples, climbing ripples, parallel and/or cross-laminated sandstone beds, convolute lamination, desiccation features, and many sole structures in the form of current grooves, prods and the biogenic sedimentary structures *Cruziana problematica*, *Rusophycus problematicum* and *Palaeophycus sp.* Some of the sandstones are erosive based and amalgamated whilst others contain shale interbeds. Loose blocks contain structures very akin to rain prints but which on close examination can be seen to result from either or both biogenic and dewatering activity. Note the consistent paleocurrent within each cycle which is basically from south to north. Note also the thin brecciated carbonate horizon towards the top of the succession which occurs in association with overbank siltstones and mudstones. Plant debris and rare crossopterygian (lobe-finned fishes) fragments may be observed in some of the sandstone beds.

Conformably overlying this basal sequence of fluvial channel and associated overbank deposits is a succession of boulder, cobble, pebble conglomerates, associated sandstones, pebbly sandstones and siltstones with rare mudstones. Although the contact is interpreted as conformable it is notable that it is marked by folded arenaceous micrite and complicated by faulting. The sequence of conglomerates is approximately 120 m thick and although the exposure is continuous, access is generally difficult. Nevertheless, a minimum of eleven fining-upward conglomerate sequences can be recognized. Each fining-upward sequence begins with coarse-grained, generally erosive based, conglomerate, which contains lenticular parallel laminated sandstone or siltstone lenses or layers and lenses of pebbly sandstone, and passes vertically into minor layers of sandstone capped by thin mudstones. Thickness of individual fining-upward sequences varies to a maximum of *c.* 20 m, with a basal sequence *c.* 3 m. The individual sequences are themselves stacked one upon the other to form a larger scale fining- upward mega-sequence, so that the whole succession, in general, fines upward.

The conglomerates are essentially matrix-supported and massive; they contain angular-subrounded clasts and are extremely immature compositionally. Mafic volcanics, granite gneiss, rhyolite, granite, granodiorite, vein quartz, chlorite schist, chert, conglomerate, and other sedimentary clasts comprise the majority of clasts. They are interpreted as the products of braided stream flows. Associated lenses and layers of sandstone, pebbly sandstone, and siltstone all possess evidence of tractional deposition and, as such, are interpreted as the products of braided stream deposits.

Mudstones capping these latter deposits were probably deposited from suspension. Thus, the entire succession of conglomerates is representative of deposition by debris flow and associated braid channels developed on an alluvial fan. The cyclic nature of individual fining-upward sequences probably reflects localized tectonic and/or climatic events within the depositional basin whereas the overall fining-upward megasequence probably reflects increased distally from source, in this case the Caledonia Uplift.

Overlying this conglomeratic megasequence is *c.* 40 m of complexly folded and faulted grey and green strata which pass upwards into fining-upward sequences of pebbly sandstones (rarely pebbly conglomerates), sandstones, siltstones, and associated mudstones similar to those observed at the start of the section. The fining-upward sequences are once again quite variable and complex but can be clearly ascertained. Of particular note in this portion of the section are the excellent examples of wave ripples, climbing ripples, detrital micas, pyrite, calcite nodules, associated phosphate, plant debris and trace fossils.

The upper (northern) portion of the section highlights the Albert Formation/Weldon Formation contact. Both the Albert and Weldon strata consist of intensely desiccated overbank mudstones and thin siltstones. The Albert Formation is, of course, by definition grey or green and the Weldon red or purplish red in colour. However, the boundary is difficult to pinpoint, as there is an obvious transition zone several tens of metres where red and grey or green sediments are interbedded. The lithologies and depositional environments do not change either within this transition zone or on either side of it. Clearly, therefore, the Albert/Weldon contact is totally transitional and no definitive and exact boundary can be ascertained.

Finally, the Boudreau section conglomerate megasequence simply represents a tongue of alluvial fan deposits which prograded into the area during late Albert Formation time.

Source

Pickerill, R.K., Carter, D.C., and St. Peter, C., 1985. The Albert Formation—Oil Shales, Lakes, Fans, and Deltas. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Fredericton '85, Excursion 6, __ p.

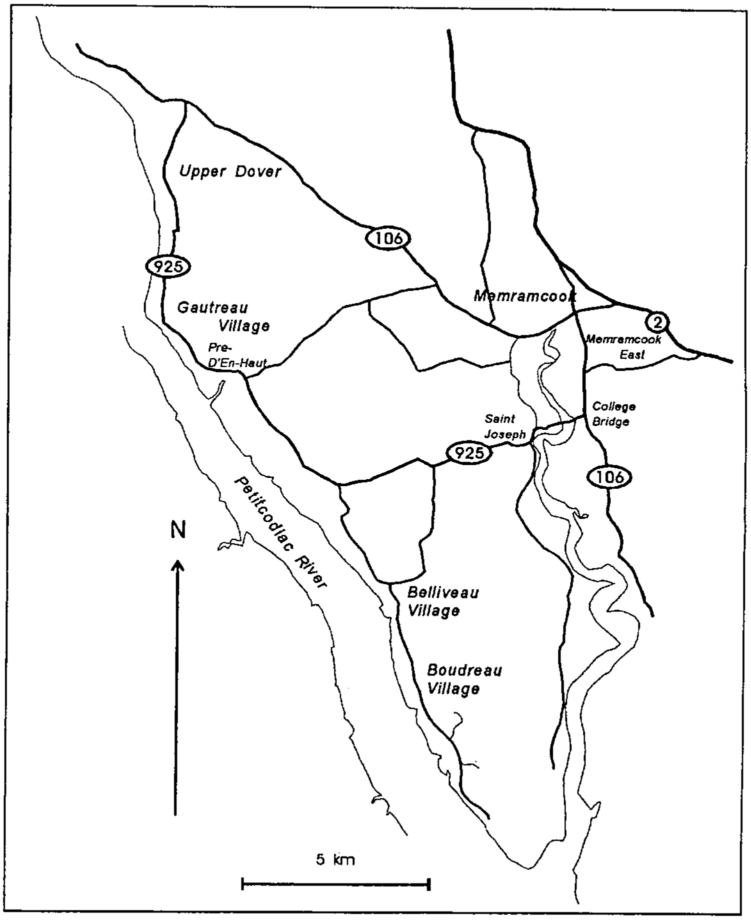


Figure 1: Location map showing Boudreau Village

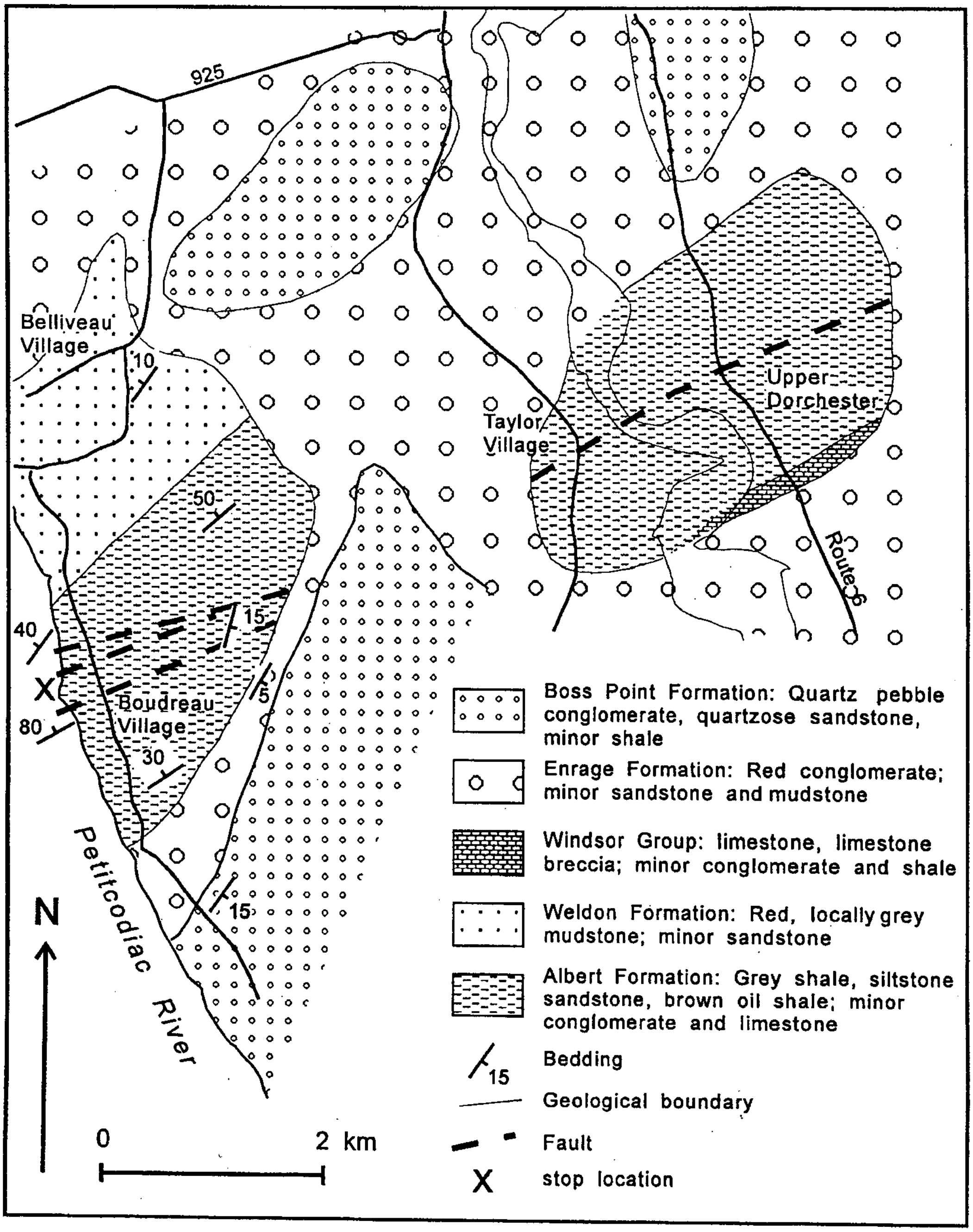


Figure 2: Generalized geology map of the Boudreau Village area. From Pickerill et al., 1985; modified by C. St. Peter, 1996.

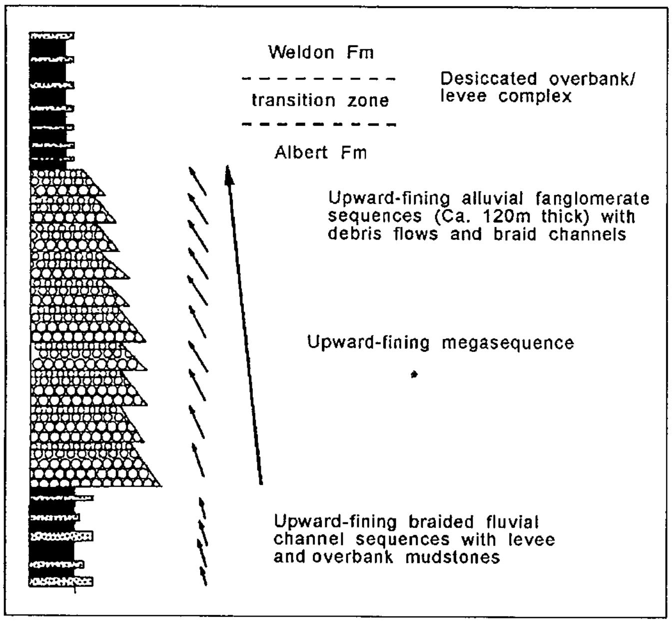
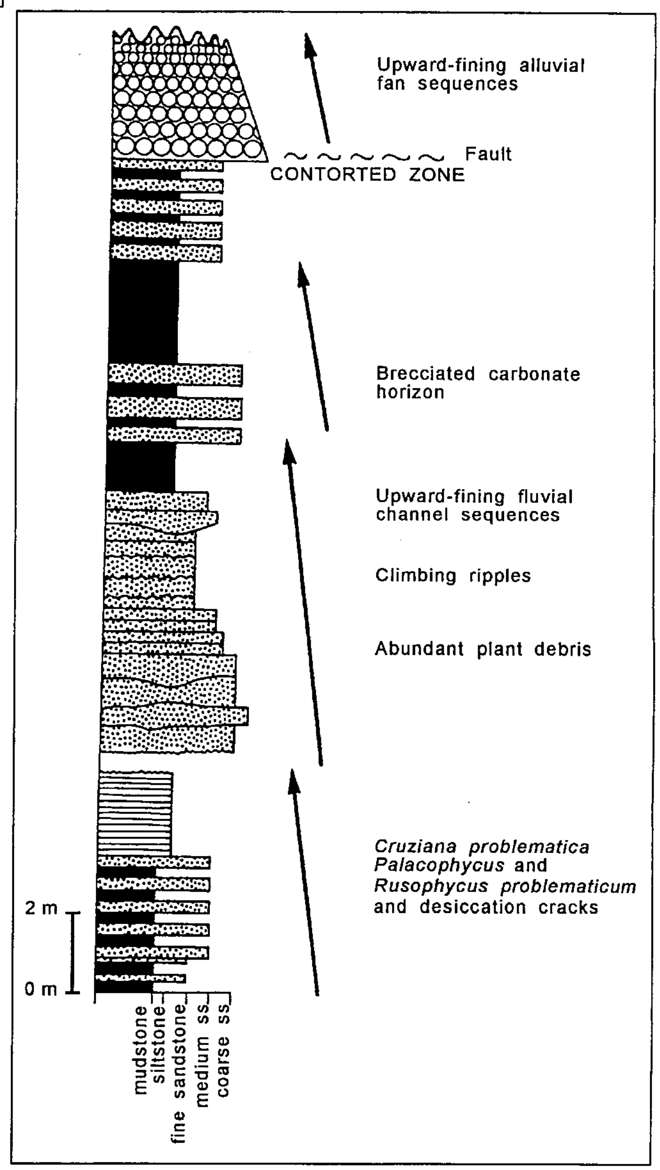


Figure 3 (above): Schematic stratigraphic section of the rocks exposed at Boudreau Village along the bankdsof the Petitcodiac River. No scale implied. From Pickerill et al., 1985

Figure 4 (right): Detailed stratigraphic section of the lower portion of the Albert Formation exposed on the east bank of the Petitcodiac River at Boudreau Village. From Pickerill et al., 1985.



Frederick Brook

Oil Shales of the Albert Formation

Highlights

- oil shales (some of which may burn if heated)
- fossil fish and fish scales

Directions

Drive 21 km south of Moncton along Route 114 to the town of Hillsborough (Fig. 1). Continue on Route 114 for 4 km until you come to an unnumbered road, designated the Albert Mines Road, going west. Proceed west on this road for 3.6 km nearly to the bottom of a large hill where there is a small church and a secondary road leading to the right (northwest). Follow this unnumbered road for about 1.8 km to a shale pile on the right (west) side of the road. The pile lies near the northeast end of the underground albertite vein and is indicated as stop 6 in Fig. 2. Note that the final 700 m of road leading to the site is not a hard surface.

Precautions

This is a remote site. Do not wander into the bush without a compass and safety pack. The road is not paved so drive with care and caution.

Significance

The area at Frederick Brook, and Albert County in general, became famous in 1849 following discovery of a unique natural substance that became known as albertite. Albertite is a black, glassy, solid bitumen that occurs as a vein along Frederick Brook (Fig. 2). The discovery led to the opening of an underground mine that began production in 1860. By 1879, the mine was depleted: a total of 200,000 tons of albertite had been exported to the United States for the production of kerosene oil and gas.

The Albert Formation underlies the Moncton Subbasin (Fig. 1); it is composed of three members. In ascending order, they are: the Dawson Settlement Member, the Frederick Brook Member, and the Hiram Brook Member. The oil shales at Frederick Brook constitute the type section of the medial Frederick Brook Member. The shales represent a spectacular example of thinly laminated strata deposited in a fresh-water lacustrine environment in Early Carboniferous (Tournaisian) time. Some shale beds contain well-preserved complete specimens of Palaeoniscid fishes and abundant fish scales.

The Albert shales are highly deformed and folded into upright, south-plunging mesoscopic folds. In addition to these larger folds, there are smaller scale recumbant folds present in exposures along Frederick Brook. The Frederick Brook rocks are bounded on the north and south by regional east-northeast-trending normal faults. To the southeast, the deformed Albert strata are unconformably overlain by the post-tectonic Late Mississippian or Pennsylvanian Enrage Formation.

Regional Geology

The Albert Formation underlies an area of about 3,000 km^2 in the Moncton Subbasin of southeastern New Brunswick. The Albert strata constitute the medial formation of the Upper Devonian to Lower Mississippian Horton Group (Fig. 2). The Horton rocks represent the first sediments to accumulate in the Moncton Sub-basin. The Sub-basin formed in late Devonian time as a result of normal faulting or strike-slip faulting of deformed Middle Devonian and older crystalline rocks. The Albert Formation is conformably underlain and overlain by the Memramcook and Weldon Formations, respectively. Both the Memramcook and Weldon Formations are composed of alluvial fan and fluviatile red beds. The Horton group rocks are exposed along the south flank of the Subbasin, where in most places the group is in fault contact with the crystalline basement of the Caledonia Uplift. On the north side of the Moncton Subbasin, the Horton rocks are exposed in a northeast-trending belt where they are typically unconformably overlain by Upper Mississippian- or Pennsylvanian-age strata.

The Horton rocks are overlain by Visean-age red beds, carbonates, sulfates, and evaporites of the Windsor Group. The Windsor represents the only known marine sequence in the Carboniferous of New Brunswick. Economically important potash beds occur in the upper evaporites of the Windsor in the Sussex area.

The Windsor strata in the Moncton Subbasin are conformably overlain by the late Mississippian to Early Pennsylvanian Hopewell Group. Along the southern margin of the subbasin, the Hopewell typically comprises coarse-grained alluvial fan deposits. In the axial subsurface part of the subbasin, the Hopewell strata are mostly fine-grained red beds lying concordantly on Windsor evaporites.

Pennsylvanian-age, predominantly gray fluviatile beds of the Cumberland Group overlie the Hopewell Group within the Moncton Subbasin. In numerous places, the Cumberland strata overstep older Carboniferous rocks along the southern part of the Subbasin and lie atop the Caledonia basement.

The Carboniferous rocks of New Brunswick have received the attention of many geologists since the early writings of Dawson in the 1860s. None of the rocks has been more inspected or has raised more interest than the Albert Formation: this is particularly true of the oil shales at Frederick Brook. The medial Frederick Brook Member of the Albert Formation in the southeastern part of the Moncton Subbasin in underlain by the Dawson Settlement Member and overlain by the Hiram Brook Member. The two latter members comprise an assemblage of alluvial and fluvial interbedded gray shale, limestone, and dolostone.

Description

The medial Frederick Brook Member is best exposed at the type section along Frederick Brook (Fig. 2). The type section comprises those rocks exposed along the brook on the north side of the road from Stop 2, down the brook to the northwest-trending fault, about 200 m north of Stop 6. The rocks in the type section are brown-weathering, fissile, laminated, dolomitic, kerogenous siltstone, and rare beds of brown massive dolostone and fine-grained sandstone.

The characteristic oil shales of the formation are best seen at Stop 1 in a shale pit in an abandoned field on the east side of Frederick Brook. The tailings from this pit are stockpiled just north of the pit. The stockpile can be easily seen across the field for 200m when approaching the site along the road immediately to the east. The oil shales in the pit are among the best grade in Canada. When split into layers, the shale can be ignited with a match. The beds average 100 litres of shale oil per ton of rock over a stratigraphc interval of 25 m.

The shales in the pit and those exposed at Stop 2, midway up the east bank of Frederick Brook, and immediately north of the road are known to containe complete fossil fish that are ascribed to *Rhadinichthys alberti* (Dawson 1868, Lambe 1909, 1910). *R. alberti* is a rather small fish averaging about 8.5 cm in length (Fig. 3). It has well-developed fins and diamond shaped scales. Clusters of black vitreous scales, ranging from 2 to 4 mm in length, are common on bedding planes.

The Frederick Brook section has been assigned an Early Carboniferous (Tournaisian) age based on miospores which imply a fresh-water origin for the oil shales. The kerogen laminations in the oil shales are probably of planktonic algal origin. The excellent preservation of *R. alberti* suggests that the lake was eutrophic, and had restricted circulation and few bottom scavengers.

The rocks along Frederick Brook reveal a complex folding history. Many of the more organic-rich oil shale beds display convoluted and chaotic synsedimentary slump folds that range from microscopic to a metre or more in wavelength. Superimposed on these sedimentary folds are two different types of tectonic folds. Tectonic folds of the first type are upright, open to tight, symmetrical to slightly asymmetrical, and south plunging, and they range in wavelength from less than 1 m to more than 200 m. Larger scale structures of this style were mapped in tunnels associated with the nineteenth-century albertite mine. A small-scale fold of this type can be seen at water level in Frederick Brook about 25 m south of the road at Stop 3. This locality can be reached by walking south along the brook about 40 m to the outcrop exposed above the water line on the east bank.

The second style of tectonic fold is recumbant, tight, and asymmetrical, with wavelengths ranging up to 2 m. A fold of this kind is exposed above the water level on the south side of a pool, at the bottom of a small waterfall at Stop 4. This fold has an axial surface cleavage (strike 085°, dip 30° south) in the hinge that dies out toward the limbs of the structure. The relative ages of the upright and recumbant tectonic folds are unknown because interference fold structures have not been observed.

The folded shales at Frederick Brook are in fault-contact to the north and south with conglomerates and mudstones (Fig. 2). Both bounding faults are egional structures that have been traced for at least 40 km. Where the traces of these structures are seen at the surface, the faults are marked by steeply dipping shear zones.

The Albert Formation is in fault-contact with a high-standing ridge of Precambrian volcanic rocks to the west. A borehole in the Albert rocks just east of the ridge reveals the presence of thick alluvial fan polymict conglomerates within the lacustrine oil shale sequence. This stratigraphic interlaying implies that the basement-bounding fault is a normal structure with a pronounced offset. The rising volcanic ridge provided coarse alluvial fan detritus directly into the adjacent lake.

The tectonic events affecting the Carboniferous rocks in the Moncton Subbasin are ascribed to the post-Acadian Maritime Disturbance. The complex array of sedimentary and tectonic structures associated with this disturbance are well represented at Frederick Brook. The major deformational events were largely concluded by late Mississippian time. This is evidenced to the southeast, where the Albert rocks and their attendant structures are unconformably overlain by the subhorizontal Pennsylvanian Enrage and Boss Point formations of the Cumberland Group.

Frederick Brook is most popularly known as the historical site of Albert Mines. A vein of solid bitumen—albertite—was mined from the area in the mid-nineteenth century. All that remains as evidence of the old mine are two shale piles at Stops 5 and 6. Samples of black vitreous albertite are still numerous and easily collected from the shale piles.

Source

St Peter, C., 1987. Oil Shales of the Albert Formation at Frederick Brook in southeastern New Brunswick. Geological Society of America Centennial Field Guide--Northeastern Section, 1987, pp. 395-398.

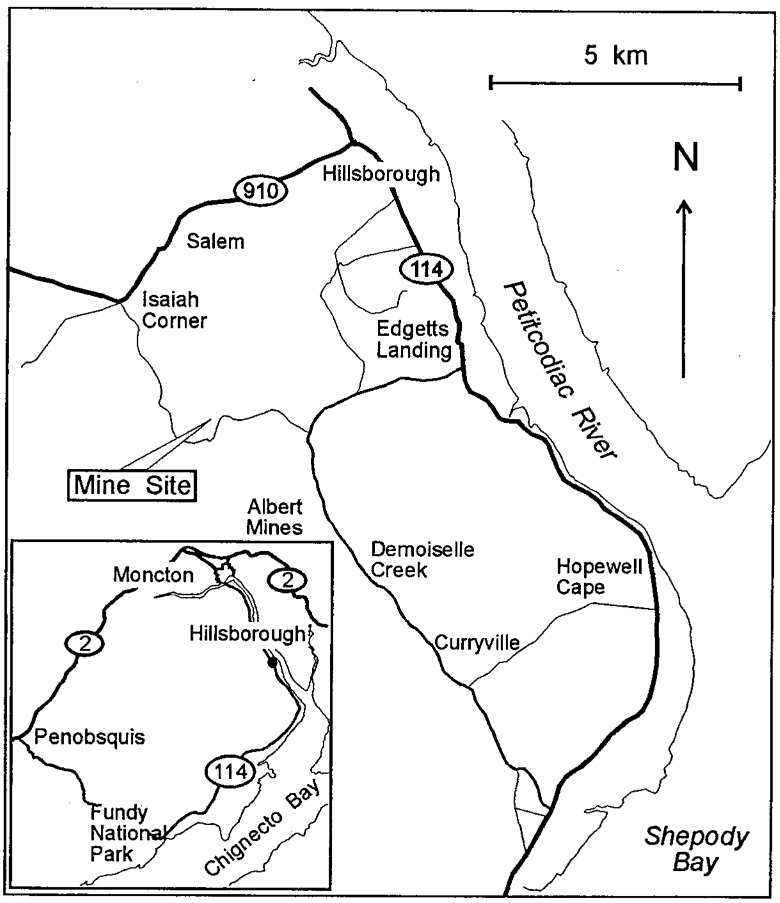
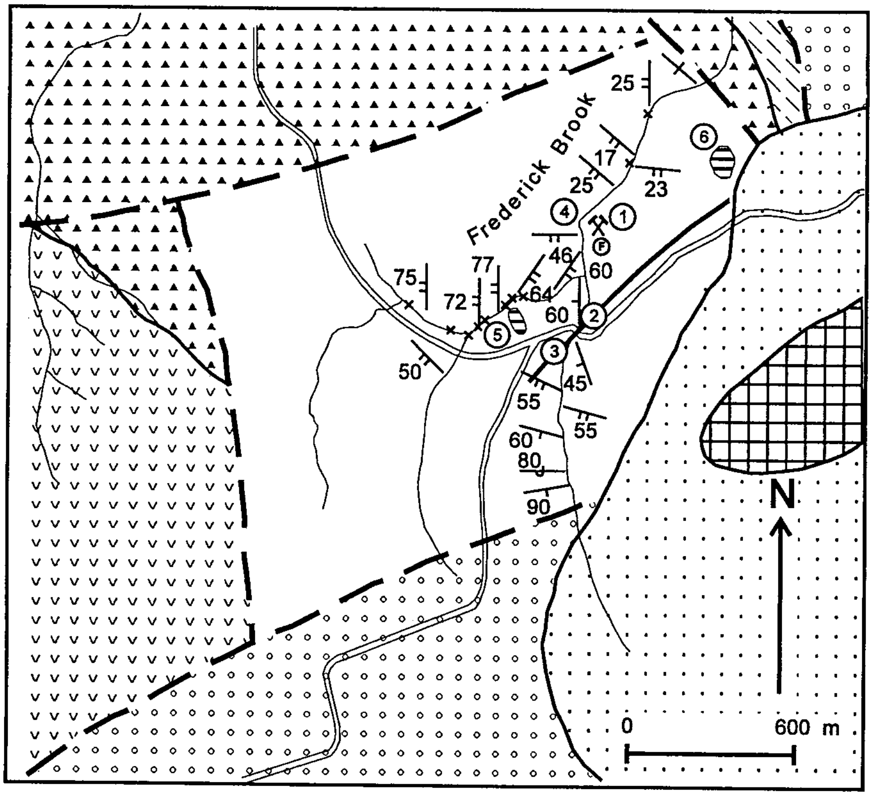


Figure 1: Location map for Frederick Brook, NB.



PENNSYLVANIAN

BOSS POINT FORMATION *Quartz pebble conglomerate, quartose sandstone*

ENRAGE FORMATION *Polymict conglomerate, sandstone and shale*

MISSISSIPPIAN

HILLSBOROUGH FORMATION *Polymict conglomerate; minor sandstone and mudstone*

WELDON FORMATION *Red mudstone and shale; minor sandstone*

ALBERT FORMATION *Polymict grey conglomerate; minor lithic sandstone, mudstone and oil shale — Oil shale, siltstone; minor dolostone, limestone, sandstone and conglomerate*

PRECAMBRIAN

COLDBROOK GROUP *Felsic and minor mafic volcanic rocks*

80 60 90 *Bedding: facing direction known (inclined, overturned, vertical)*

55 *Bedding: tops unknown (inclined)*

Geological boundary

Fault

× *Outcrop*

Location of underground albertite vein

Ⓕ *Fossil locality*

③ *Stop locality*

Figure 2: Geological map of the Frederick Brook area. Modified from St. Peter, 1987.

Hopewell Cape

Stratigraphy of the Windsor and Hopewell Groups

Highlights

Stromatolitic and oncolitic, sparsely fossiliferous limestones and siliciclastics of the Windsor Group overlain by coarse, pebble to boulder, conglomerates and sandstones of the Hopewell Group.

Directions

At Moncton cross over to Riverview and go south on Route 114 through Stoney Creek and Hillsborough (Fig. 1). At Hopewell Cape turn into the Provincial Park and go down to the beach. Walk south along the beach among the flowerpots to the furthest south outcrop which is the Windsor Group.

Precautions

The field trip involves a coastal section in Shepody Bay (Bay of Fundy) which has some of the highest tides in the world. Although the tide does not pose a serious problem at this stop, the water level does rise rapidly and caution must be exercised at all times. The section is best viewed when the tide is out, so use tide tables to calculate the times when to look at the rocks. The cliffs are loose and unstable so do not venture too close. As a guide, do not go closer on the beach than the height of the cliff. Do not climb any cliff.

Introduction

At Hopewell Cape dark grey fossiliferous limestone of the Windsor Group, subzone B, underlies coarse red conglomerate of the Hopewell Group (Fig. 2). The succession dips to the north-northeast at approximately 25-40°. In contrast to playa type silt and fine-grained sandstone, which is characteristic of the Hopewell Group elsewhere in the Shepody Bay region, the Hopewell Cape sequence is mainly coarse, poorly sorted conglomerates and arkosic sandstone in which sub-angular to sub-rounded clasts of granite, gneiss, and volcanic rocks predominate. Clasts range in size from 10 to 15 cm in diameter, rarely exceeding 30 cm and the beds are defined by clast size and average 20 to 30 cm thick with a few up to 100 cm. Most beds are massive and structureless. The succession is interpreted to represent a high-energy fan-head or mid-fan deposit of an alluvial fan spreading easterly from a nearby source area in the Caledonia Mountains in the wake of a retreating Windsor sea.

The Hopewell Group has a series of widely spaced joints that are deeply weathered and this gives rise to the "flowerpots" that dot the shore.

Description of the Windsor Group (from McCutcheon, 1981)

The Windsor Group sediments at Hopewell Cape are the reference section for the Damoiselle Creek beds and are divisible into three parts, two thin grey units separated by a much thicker red bed unit. Overall the thickness of the sequence is in excess of 25 m but it is variable as shown by the two sections of figure 3.

In the western part of the outcrop (Fig. 3) the lower grey unit is 3.2 m thick and begins as a yellowish green to dark greenish grey, thinly bedded, calcareous mudstone (maximum grain size 0.05 mm) in gradational contact with the underlying greyish red mudstone of the "Windsor B" beds. The mudstone is in sharp contact with an overlying pale brown, medium-grained, calcareous sandstone, which is in turn followed by a brownish to greenish grey to black fossiliferous wackestone with a few thin sandstone interbeds. Fossils, which are not abundant, include pelmatozoan fragments, unidentified shelly material, ostracods, a high-spired gastropod, and scarce fragments of *Paleocrisidia* sp. Intraclasts occur locally toward the top of the bed and throughout are very thin discontinuous iron oxide laminae. The grey unit is capped by an olive grey medium- to coarse-grained calcareous sandstone and/or very siliciclastic grainstone that is erosional into the underlying wackestone. The upper contact is gradational with the overlying red unit, where the sediments are fine- to medium-grained, thin to medium bedded sandstone and mudstone.

In the intertidal area to the east the upper grey beds are thicker and contain stromatolites—both columnar forms and large "heads". Inside some stromatolites the maximum grain size is 0.1 mm and there are no oolites whereas on the outside the maximum grain size is 2.0 mm and surficial oolites are numerous but small (0.5 mm). Other stromatolites contain surficial oolites and are only distinguished from the surrounding rocks by their higher carbonate content. The sandstone surrounding the stromatolites is mainly siliciclastic but carbonate grains (max. size 0.3 mm and max. 5 per cent) and surficial oolites (15–50 per cent) are also present. Siliciclastic clasts are subangular to subrounded metamorphic rock fragments, quartz, and feldspar in a 30 per cent carbonate matrix.

The middle red unit bed in the eastern section is about 14 m thick and consists of greyish red non-calcareous, conglomerate and sandstone, plus one bed of siliciclastic grainstone and/or calcareous sandstone. The conglomerate beds exhibit characteristics typical of sheet-flood (pebble imbrication and cross-bedding) rather than debris flow deposits. The pebbles and cobbles are lithologically similar to metamorphic rocks to the west in the Caledonia area.

The upper grey bed unit gradationally overlies the red unit, contains numerous alternating lithologies, and is about 8.6 m thick where measured. It is unlike the lower grey bed unit in that oncolites are abundant in the siliciclastic wackestone beds, but in many other respects the beds are similar. It consists of various shades of grey to green to brown to black sandstone to siliciclastic wackestone and grainstone with or without stomatolites and sparse fossils. Algal patches occur in several beds and may be in life position or transported and surficial ooliths are present, many with cores of these algal intraclasts. Some beds contain an abundance of quartzo-feldspathic silt among the thrombolitic patches of algae. The upper contact of the Damoiselle Creek beds is placed at the top of the grey unit. The contact is sharp but it does not exactly mark the grey to red colour transition. The upper half of the topmost siliciclastic grainstone is greyish red rather than greenish grey.

Miospores, recovered from a thin mudstone unit stratigraphically above these beds (Fig. 3), indicate a Lower Windsor age (middle Visean) for these rocks.

Source

McCutcheon, S.R., 1981. Stratigraphy and Paleogeography of the Windsor Group in Southern New Brunswick. New Brunswick Department of Natural Resources, Mineral Resources Branch, Open File Report 81-31, 210 p.

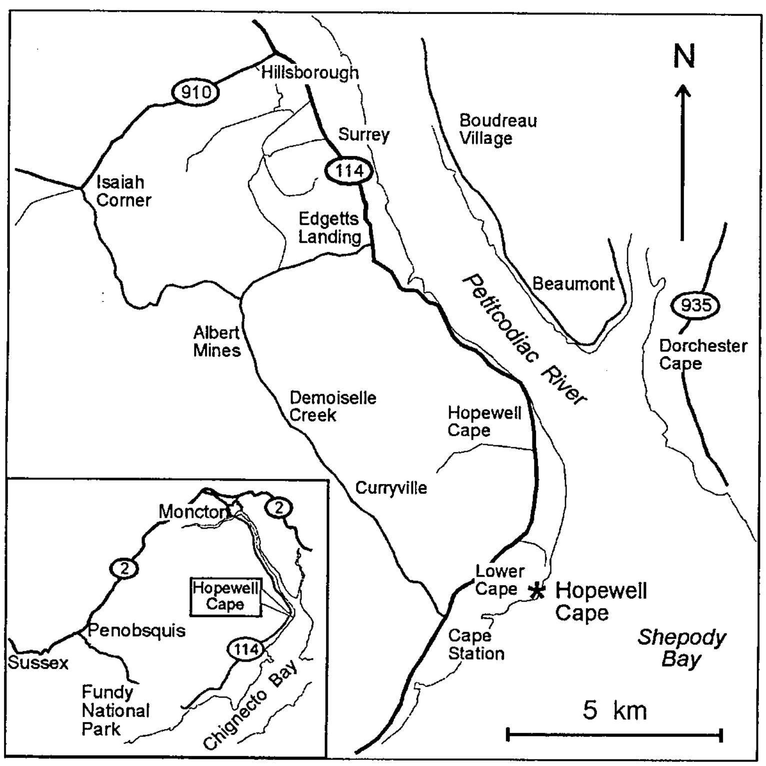


Figure 1: Location map showing Hopewell Cape, NB.

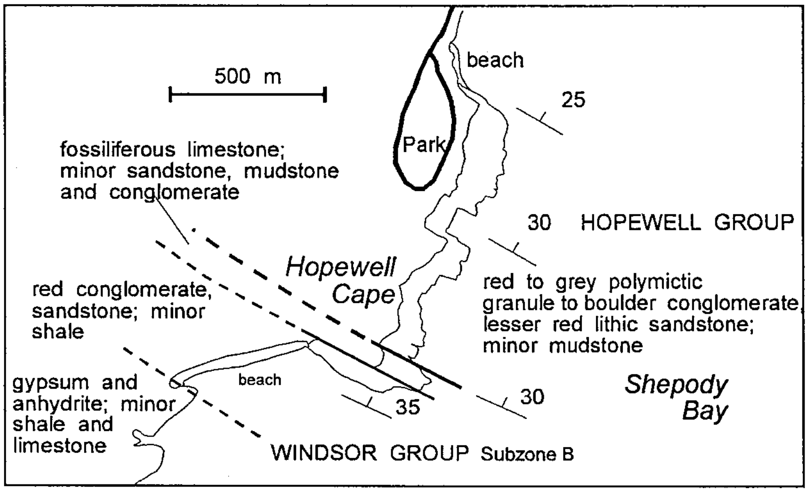


Figure 2: Geological map of Hopewell Cape. Modified Rast (1973) and St. Peter (1996)

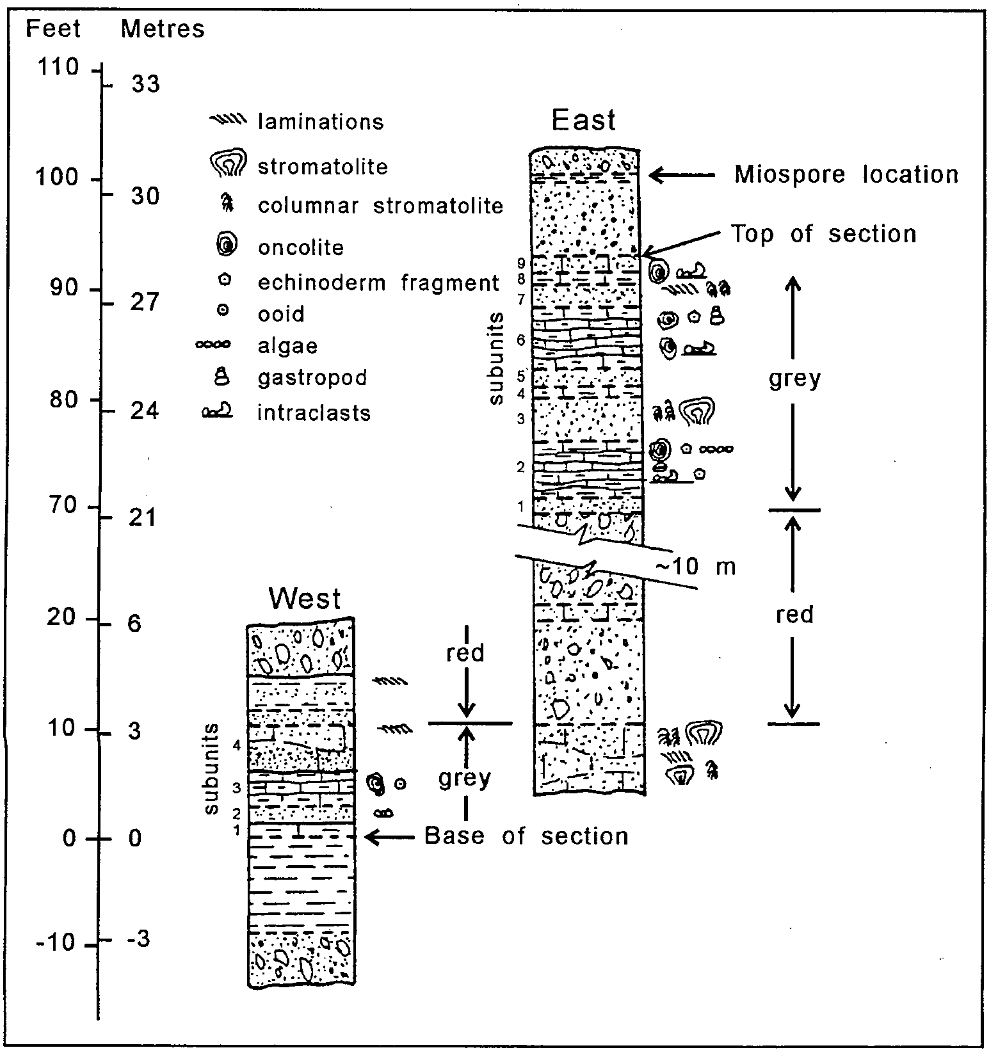


Figure 3: Graphic logs of the Windsor B rocks at Hopewell Cape.
Modified from McCutcheon, 1981.

Taylors Island

Directions

The Irving Nature Park is located on Taylors Island off Highway 1 in the city of Saint John (Fig. 1).

A driving road and a hiking trail around the island provide easy access to shoreline outcrops.

Highlights

The rocks were formed by volcanic eruptions at least 300 million years ago. Basaltic lava flows are interbedded with mudrocks which display soft-sediment deformational structures caused by slumping. In places the lava flows are mixed with the mudrocks, forming pillow-like forms. In some places, the lava intruded underground along layers in the mud to form sills. The surfaces of outcrops along the shoreline have been sculpted into wave-like forms and scratched by rocks embedded in glaciers which moved over the area a few tens of thousands of years ago. The ice melted away about 13,000 years ago, depositing piles of mud, sand, and gravel that are now eroding to form the beaches around the island.

Introduction

Taylors Island is mainly composed of dark grey volcanic rock called basalt, interlayered with red or grey sedimentary rocks called siltstone. The basalt formed from lava flows, similar in composition to those of the present-day Hawaiian Islands. The lavas formed by melting of rocks at depths of 40 or 50 km inside the Earth to form liquid rock called magma. The hot magma rose and flowed out of cracks onto the surface of the Earth, where we call it lava. Because of the chemical composition of the lava (high in silicon, oxygen, magnesium, iron, calcium, and sodium), the minerals that crystallized as the lava cooled were mostly plagioclase, pyroxene, and magnetite. However, because the lava cooled quickly at the surface of the Earth, the minerals are very small and difficult to see when you look at the rock.

After the lava cooled and hardened (crystallized), mud settled on top. After a few years, another lava flow took place, and was again covered by mud. This process happened many times (Fig. 2a); some flows are thin, others are many metres thick. With burial and time, the mud hardened to become the siltstone now interlayered with the basalt flows. In some cases, the lava did not reach the surface, but flowed underground parallel to the sediment layers to form sills. Originally, the mud layers, lava flows, and sills were horizontal (approximately parallel to the Earth's surface), like the mud exposed on Saint Rest Beach when the tide goes out. However, subsequent deformation tilted the rocks into their present nearly vertical orientation (Fig. 2b). The top of the pile is toward the east, out under the Bay of Fundy.

Over most of the island, the basalt and siltstone are covered by soil, sediment, and vegetation, but the rocks form outcrops (bedrock exposures) around the shore of the island. Headlands are formed mainly of basalt, which is hard and resistant to erosion, whereas the softer more easily eroded siltstones generally underlie the coves and beach areas (Fig. 1). At the northern tip of the island, and in the area around the visitors' information centre, deposits of sediment (mud, sand, and gravel) form steep, easily eroded banks. These sediments were deposited by melting glaciers about 13,000 years ago. The time in Earth history when the underlying volcanic and sedimentary rocks formed is not known exactly but was at least 300 million years ago, and possibly as long ago as 550 million years.

Guidelines

A full morning or afternoon could easily be spent on Taylors Island, which also has many biological features of interest, and the scenery is beautiful (if the day is clear!). Access to the outcrops is best at mid- to low tide levels. At some times of the year, vehicles have to park at Stop 1. If you can drive around the island, access to the outcrops is easy and requires very little walking. Care should always be taken when clambering on rocks along the shoreline, however.

Field Trip Stops

You can find clues about the conditions at the time when the volcanic and sedimentary rocks were forming by looking at these rocks at several places around the island, starting at the parking area at the entrance to the island (Fig. 1).

Stop 1

Stop at the parking lot at the entrance to the Nature Park at the west end of Saints Rest Beach. Walk across the beach to the rock outcrop about 10 m south of the parking area. In this area, you can see the three main components of the geological story of Taylors Island.

The rock outcrop in front of you is made mostly of dark grey basalt, formed by the cooling and crystallization of lava flows. The prominent vertical band of red layered rock that crosses the outcrop is a sedimentary rock called siltstone. It originated as layers of red mud over which the lava flowed. The fine layering (laminations) in the siltstone may be the result of seasonal variations in the size of mineral grains being deposited—rainy seasons cause more run-off and deposition of larger grains, whereas slow-moving water in drier seasons carries and deposits finer grains. Some of the layered mud was trapped in the lava, to form the siltstone band. You can think of the siltstone as being like a raft floating in the lava. If you look closely at the rocks, you can see that the basalt is darker coloured close to the siltstone; this is due to its very fine mineral grain size where the hot lava "froze" up against the cold mud. You can also see some holes in the basalt left by bubbles of gas that later escaped. Some of the holes are filled with a green mineral called epidote, or the white minerals quartz (SiO_2) and calcite ($CaCO_3$). The siltstone is red because the mud was exposed to air at the time of deposition and iron became oxidized to red minerals hematite (Fe_2O_3) and limonite (FeOOH).

The subhorizontal scratches and wave-like features that you see on the surface of the outcrop formed much more recently. If you look carefully, you can see that there are two sets of scratches, not quite parallel. These scratches were made by glaciers moving across these rocks. The ice itself did not make the scratches because ice is not hard enough to scratch rock. But lots of rocks were imbedded in the ice and they carved the scratches. The weight of ice on these imbedded rocks (the glacier was probably several kilometres thick) ground away the underlying bedrock and carved scratches and the wave-like appearance of the outcrop.

The other geological feature that you can see here is the big pile of mud, sand, and gravel—the unconsolidated material that sits on the outcrop. This was deposited by the glaciers as they melted. This material formed a big dam which forced the Saint John River to carve in its present course through Reversing Falls. These unconsolidated deposits are easily eroded—you can see how they are slumping down the cliff—but much of the coarser sediment is quickly deposited again to form and maintain the beach behind you!

Stop 2

Drive or hike around the island to Stop 2 on the west coast of the island. If you are driving, a parking area is located at this stop.

In this area, the basalt magma did not reach the surface to form a lava flow like we saw at Stop 1, but instead spread out as thin layers called sills in a pile of sediments. These sediments were buried and hardened when the magma squeezed between the layers, cooled and crystallized, so mud did not mix with the magma (in contrast to what you will see at Stop 4). The minerals in the sills are bigger than in the lava flows at Stops 1, 3, and 4 because the magma in the sills cooled more slowly and crystals had more time to grow.

Stop 3

Continue driving or hiking to the Gorge Rocks on the south coast of the island. If you are driving, a parking area is located near this stop.

Basaltic lava flows in this area contain many veins and pods containing a green mineral called epidote and the white minerals quartz (SiO_2) and calcite ($CaCO_3$). These minerals are called "secondary" because they were added to cracks and cavities in the rock after it had crystallized. Economic minerals, such as gold, often occur with these secondary minerals.

If you walk to the west side of the beach west of the Gorge Rocks, you will see examples of "soft-sediment deformation." This happens when mud becomes unstable and slumps and slides down sloping surfaces after it is deposited. The laminations in the sediment become crumpled and contorted. Slumping may have been caused by shaking of the Earth during an earthquake.

Stop 4

Continue driving or walking to Stop 4 on the southeast coast of the island. If you are driving, go past the stop to the next parking area, and walk back about 100 m to the outcrop.

This outcrop contains wonderful examples of a structure called peperite. Imagine a crack opening in the Earth. Lava pours out of the crack and covers a muddy beach. In places the lava flows into the water, or mixes with cold mud. In such places, the lava forms bulbous, blobby shapes—the resulting mixtures of basalt and siltstone are called peperite, and they are mainly at the bottoms of the lava flows. In places in the basalt you can see holes left by bubbles of gas that later escaped. Some of these holes were later filled by epidote (green), quartz, or calcite.

Contributors

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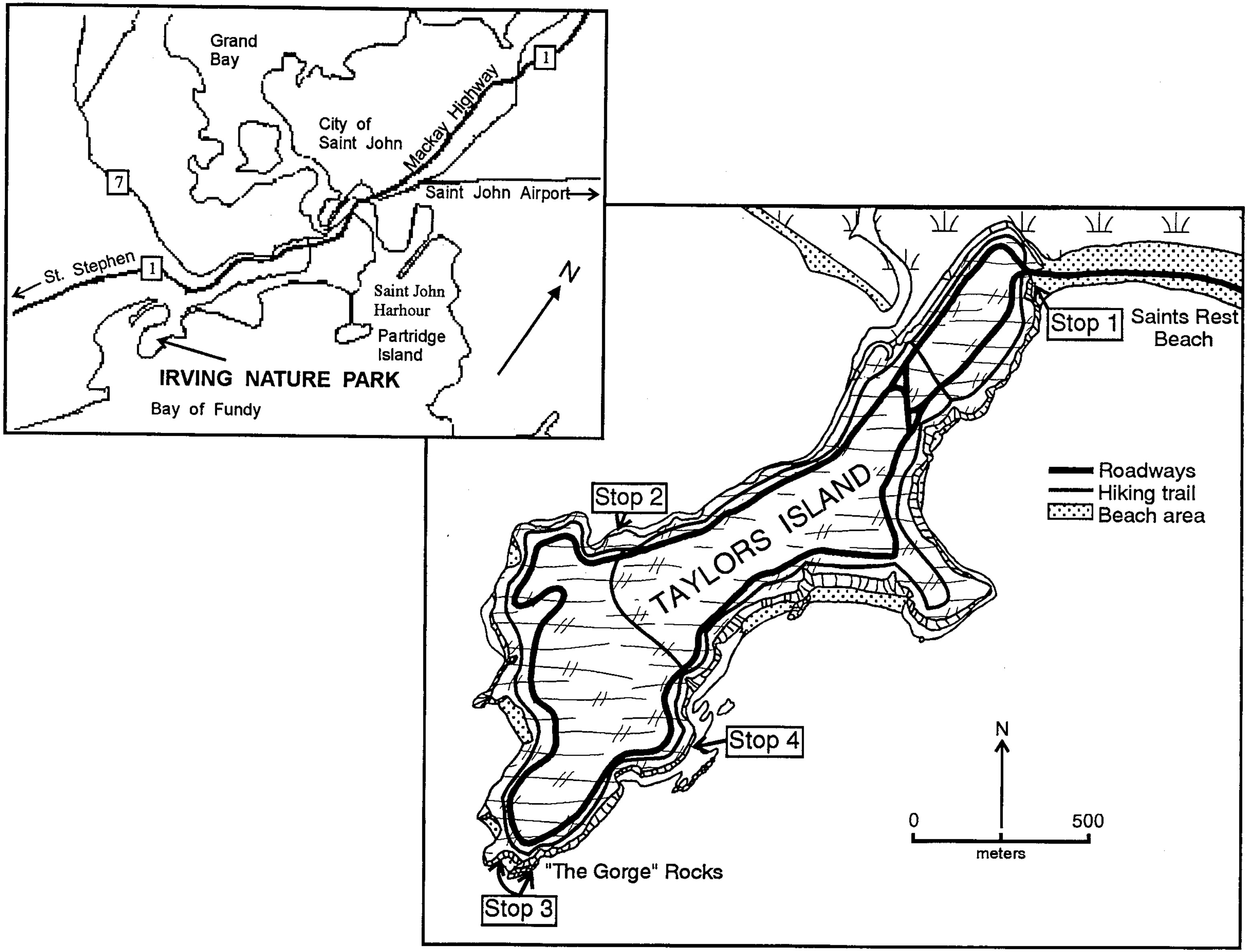


Figure 1: Sketch map of Taylors Island, showing the locations of the stops

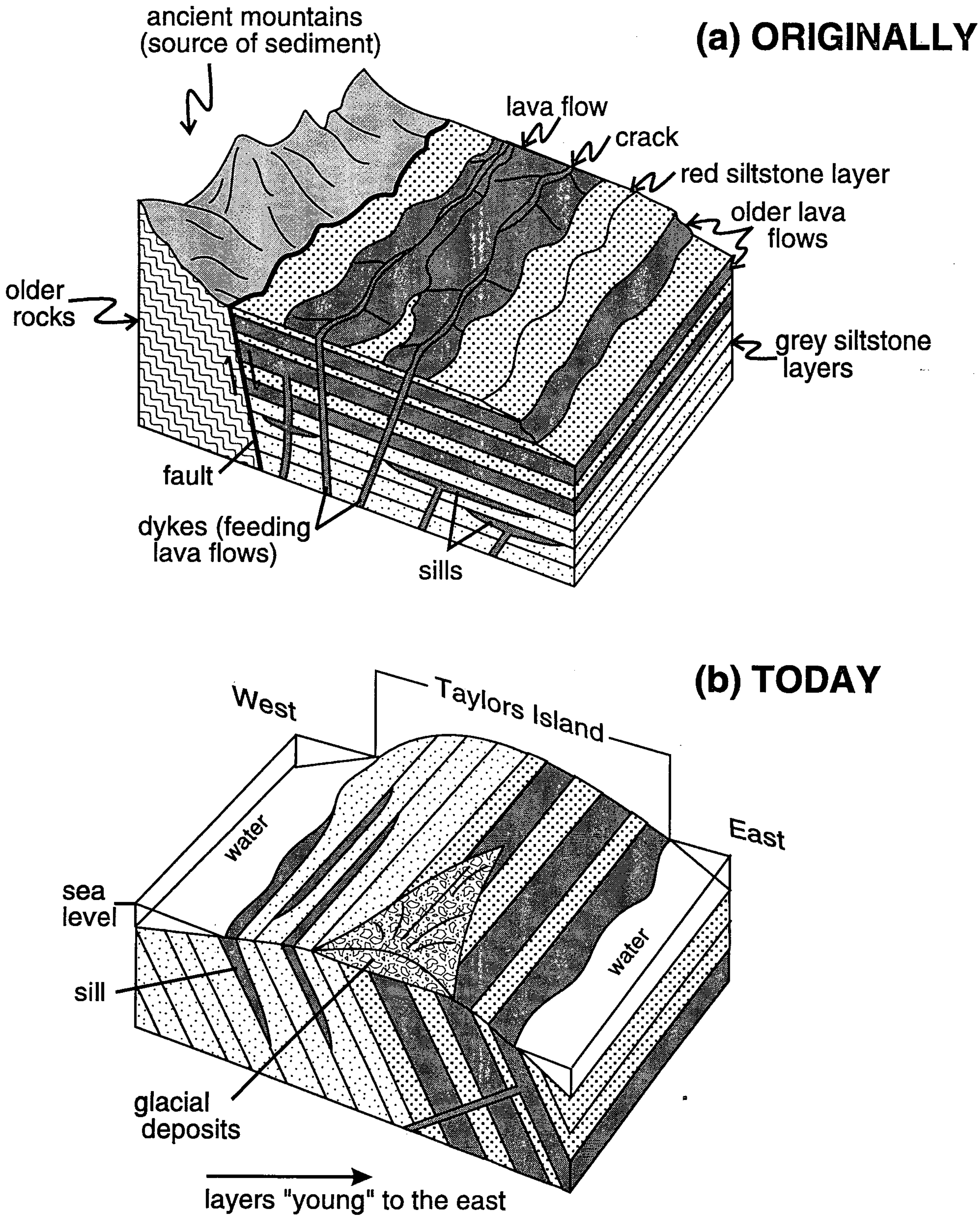


Figure 2: Block diagrams to illustrate some of the geological history of Taylors Island.

St Andrews

Stratigraphy and associated igneous rocks of the Upper Devonian Perry Formation

Purpose/Highlights

- Triassic diabase dyke
- Upper Devonian mafic lava flows
- Upper Devonian terrestrial clastic sediments

Location

The two stops are on the St. Andrews Peninsula, near the town of St. Andrews (Figure 1). The first stop is outside the Huntsman Marine Science Centre, Aquarium and Biological Station (Fig. 1). Follow signs to the Huntsman Centre and park at the south end of the large car park. Walk south through stratigraphic section to Joes Point. Stop two is on Passamaquoddy Bay accessed from the Bar Road. To get there from stop 1 drive back up the road from the Huntsman Centre and follow the paved road at the junction turning right. Cross the St. Andrews Golf Course and turn left onto Cedar Lane, driving north alongside the course. At the stop sign turn right onto route 127 towards St. John. Turn right onto the Bar Road at 0.6 km along route 127. Park on the raised area at the west end of the gravel bar to Ministers Island (Fig. 2).

Precautions

The stops are along the shore where tidal ranges are extreme and tidal currents swift. If you stick to the outcrops it is not dangerous but exercise extreme caution at all times. At Bar Road the gravel bar provides a walkway at low tide to Ministers Island but travelling on it is not advisable since it is only exposed for about 2 hours each tidal cycle. Ministers Island is a Federal Historic Park—there are tours available.

Introduction

The Upper Devonian Perry Formation has been recognized as a post-tectonic molasse, deposited during uplift following the Lower Middle Devonian Acadian Orogeny. It extends mainly on the west side of Passamaquoddy Bay between Perry, Maine, and St. Andrews, New Brunswick, with scattered outliers in the Cobscook Bay area, Maine, and in the Beaver Harbour area along the Bay of Fundy in southwest New Brunswick. The Perry Formation unconformably overlies the Lower Devonian Eastport Formation and non-conformably overlies the red Beach Granite in the St. Andrews area and adjacent coastal Maine. The unconformity is exposed in many places along the coast.

The Perry Formation consists of mainly of conglomerates and sandstones with finer grained sediments interpreted to be scree, alluvial fan, fluvial, overbank, and lacustrine deposits. The middle sequence includes three basalt lavas, two of which are discontinuous and pre-, syn(?)- and post Perry diabase dykes cut all the strata. A unique rhyolitic tuff near the base of the Perry Formation is a volcanic marker horizon of probably regional extent, linking the Perry Formation in the Perry-St. Andrews area with that in the Beaver Harbour area. Overall the strata dips gently to the southeast with local noncylindrical folds, a major sigmoidal strike swing, faults, orthogonal joints, closed spaced joints in individual conglomerate pebbles, intense fractures in massive siltstones, and planar exfoliation in sandstones; the origin of some of these structures is enigmatic.

Description

Stop 1: Joes Point, St. Andrews

The lower basalt lava (Loring Cove Basalt Member) outcrops at the southwest corner of the car park. It strikes NNE under the Biological Station buildings 150 m to the north, and dips gently ESE, at the south end of the major sigmoidal strike swing across Brandy Cove.

Conglomerate and minor interbedded pebbly sandstone in the lower half of the Brandy Cove Member outcropping along a low cliff to the east and south of the car park for about 200 m south to the middle basalt lava. Bedding dips ESE to SE between 10° and 22°. Boulders form 2 per cent of the conglomerate. Erosion gullies in massive conglomerate, 50-85 m north of the middle basalt outcrop, follow 055° joint planes; some gullies may follow minor faults.

The middle basalt lava has fewer phenocrysts than the lower and upper basalt lavas, and vesicles and amygdules are usually absent.

Conglomerate and minor sandstone beds in the upper half of Brandy Cove Member outcrop around and form Joes Point, dipping 14-21° SE. The conglomerate consist of up to 5 per cent boulders and many cobbles. Orthogonal joints in the sandstone beds trend NE and SE.

The upper basalt lava (Joes Point Basalt Member) outcrops on the coast 125 m east of Joes Point. The upper half of the basalt is vesicular and amygdaloidal.

Boulders and cobbles are common in the massive conglomerate immediately above the Joes Point basalt in map unit 5c at the base of the St. Andrews Member, 300 m east of Joes Point. At this locality, minor faults trend 040°-055°. Closely spaced vertical joints are prominent within individual cobbles and boulders (sliced up like loaves of bread); the joints, which do not penetrate the surrounding rock, strike 048°, subparallel to rare through-going joints elsewhere in the conglomerate.

Stop 2: Bar Road, St. Andrews.

Joes Point Basalt Member: The basalt lava is very fine-grained to fine-grained, with plagioclase phenocrysts and relict pyroxenes, and is very altered; vesicles vary from abundant, in places coarse, to sparse.

The syncline in the basalt plunges gently SSE. The basalt is conformable with sediments above and below at the coastal outcrop 100 m WSW of Stop 2. The basal contact of the basalt in subsurface has been located by magnetometer picks (Fig. 2) made from eight magnetic anomaly profiles across the basalt.

Brandy Cove Member: The strike of bedding changes from NNE near the Joes Point basalt to NE about 250 m west of Stop 2, and to 100° farther north along the coast. The pebbly sandstones and conglomerates (map unit 3p, Fig. 1) form the top part of the Brandy Cove Member.

St Andrews Member: The conglomerate which forms a recessive outcrop between basalt and sandstone 125 m SSW of Stop 2 is the basal unit (map unit 5c, Fig. 1) of the St Andrews Member. It is overlain by pebbly sandstone and red sandstone beds (map unit 5p, Fig. 1) which outcrop intermittently along the coast 200-400 m south of Stop 2, dipping ENE at 20-30°.

Joints occur mostly in two approximately orthogonal orientations:

055° (036-065°) strike, subvertical

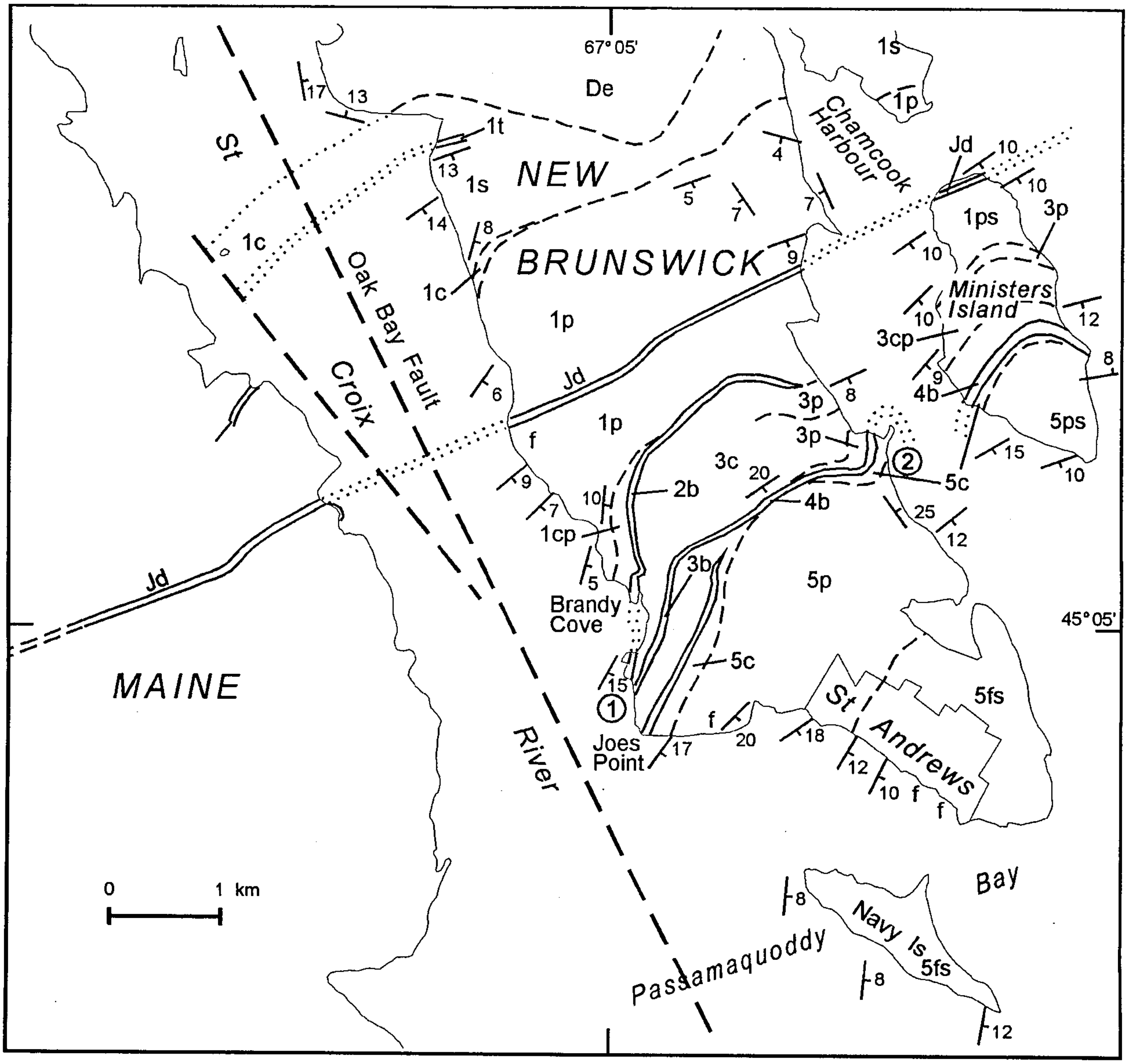
135° (125-144°) strike, subvertical

In areas of pebbly sandstone with few joints, individual igneous pebbles show closely spaced fractures subparallel to the 055° joint direction, and some apparently unfractured pebbles split persistently in the 055° direction when hammered. The 135° joints commonly terminate against the 055° joints, but in places the terminations are mutual. Sporadic exfoliation is present in the sandstones parallel to both joint directions.

The ENE-trending faults (Fig. 2) are strike-slip faults based on slickensides. Sandstone on the south side of the northern fault is faulted against conglomerate overlain by sandstone on the north side 200 m south of Stop 2, suggesting that the strike-slip movement is dextral. At the southern fault, bedding attitudes change from 160/20 E north of the fault to 044/12 E south of the fault.

Source

Stringer, P., Burke, K.B.S. and Dunn, T., 1991. Stratigraphy, Structure and Associated Igneous Rocks of the Upper Devonian Perry Formation in the St Andrews Area, Southwest New Brunswick, and Adjacent Coastal Maine, New England Intercollegiate Geological Conference Field Trip, 83rd Annual Meeting, September, 1991, Princeton, Maine.



Map Units

Jd Lower Jurassic Ministers Island diabase dyke

Dp Upper Devonian Perry Formation
- 5 St Andrews Member
- 4 Joes Point Member
- 3 Brandy Cove Member
- 2 Loring Cove Basalt Member
- 1 Lamb Cove Member

Dg Lower Devonian Red Beach Granite

De Lower Devonian Eastport Formation

Lithology of Perry Fm map units

- b basaltic lava
- c conglomerate
- f fine-grained sandstone
- p pebbly sandstone and sandstone
- s brick red siltstone
- t rhyolitic ashfall tuff

Symbols

12 bedding, inclined

contact (approximate: igneous, sedimentary)

contact offshore

fault

② stop location

Figure 1: Location map showing regional geology. From Stringer et al., 1991.

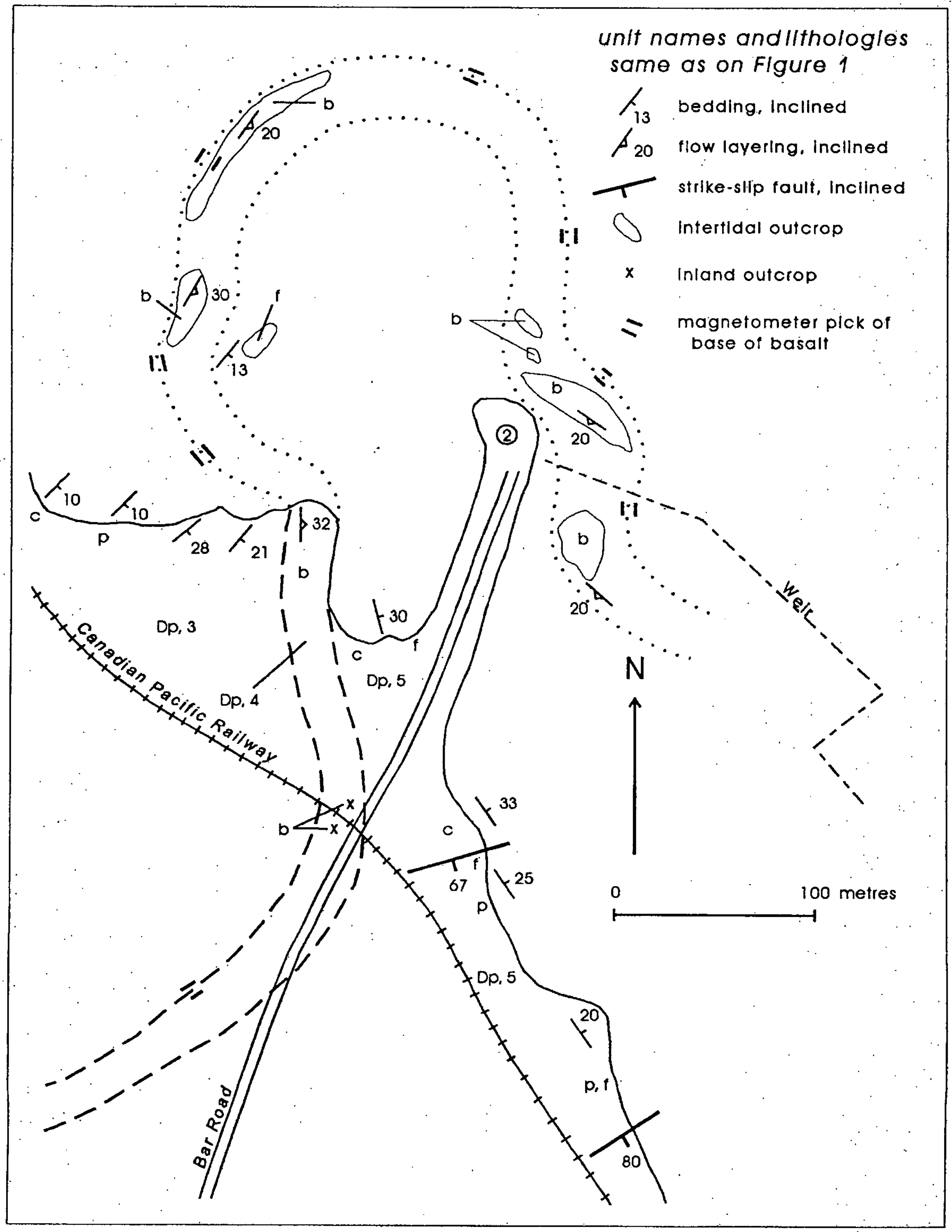


Figure 2: Detailed geology map of Bar Road, St. Andrews. Modified from Stringer et al., 1991.

Grand Manan Island

Polydeformation in the pre-Triassic Rocks

Location

Grand Manan Island, in the Bay of Fundy, is accessible by ferry from Blacks Harbour on the mainland. Check with ferry crossing times before leaving. There is no charge to go to the island, only to leave it. The ferry-boat takes about 1½ hours and passes through some of the most densely populated areas of aquatic mammals in the Bay of Fundy. Take your binoculars to view whales, porpoises, and dolphins. If you are staying overnight on the island there are facilities for camping, as well as rooms in motels.

There are two stops described in this guide, one at the Thoroughfare and the other at Ingalls Head (Fig. 1). After disembarking from the ferry take the main road, Route 776, south towards Grand Harbour. The Thoroughfare is accessed at the northern end of Grand Harbour village by a paved road, then gravel, that goes southeast at the first major bend. There is a fish/lobster processing plant at the end of the gravel road as well as a lobster pound in the waters of The Thoroughfare. At low tide the road continues to Ross Island across the passage, and I do not recommend you travel on it. The outcrops to examine are on both sides of the channel from the plant to the cages.

The second stop starts at the ferry terminal to White Head Island at Ingalls Head. At the southern end of Grand Harbour village take the paved road to the ferry terminal and park at the south side of the jetty to avoid mix ups with the vehicular line-up for the ferry. The rocks to examine are south along the coast for about 700m.

Precautions

The exposures are coastal so awareness of the tides is necessary. Although the tidal range is not great, the tidal currents are tremendous, putting up standing waves a metre or more high and speeds of 5 to 7 knots. Do not cross over to Ross Island without knowing how you are getting back, and wear rubber boots in the wet muddy intertidal regions. Stay off the tops of the small cliffs because they are unstable: examine the rocks from below. The outcrops break or split along closely spaced partings in the rock giving sharp edges and in combination with wet seaweeds in the intertidal regions are extremely slippery and dangerous.

Both stops have you looking at rocks on private property and walking around business operations. Please be respectful and mindful of the people who live and work there, do not block vehicles, roads, or doorways.

Introduction

The western and central portions of Grand Manan Island are underlain by Triassic-aged volcanics and sediments, whereas the eastern side and offshore islands are underlain mainly by polydeformed sedimentary and volcanic rocks; intrusives, deformed and undeformed, are found at North Head and Three Islands respectively (see Fig. 1). The ages of the older rocks are unknown; they are believed to be late Precambrian to Ordovician-Silurian, based on correlation of structural styles and metamorphic ages to rocks on the shore of mainland New Brunswick.

The oldest rocks, found in the Thoroughfare and on Ross Island (Fig. 2), consist of interbedded black pelite and quartzite. Quartzite clasts within volcaniclastic sequences on Nantucket Island and Whitehead Harbour suggest the overlying volcanic rocks are unconformably overlying these sediments although the structural data suggests the angularity between the two is not appreciably angular. The volcanics consists of flows, pillowed lavas, hyaloclastics, and volcaniclastic sandstones and siltstones with appreciable amounts of non-volcanic source rocks (i.e., quartzite, limestone, granite and gneiss).

The deformation in the sedimentary and volcanic rocks is variable and poly-phase. The primary layering is a transposed bedding (S_0 parallel to S_1) with later deformations cutting this fabric. Minor F_2 folds can be seen but no major folds have been identified and the S_2 crenulation cleavage is fairly constant throughout suggesting an absence of later folds as well. The best examples of these structures can be found in the pelitic rocks and Stringer and Pajari (1981) have been able to identify at least five phases of deformation.

The initial deformational phase is characterised by an alignment of the mineral grains into a penetrative fabric subparallel to bedding in the sedimentary rock but it is variable in the volcanic rock from absent to a spaced platy cleavage, or subparallel alignment of elongate volcaniclastic fragments. The S_2 cleavage varies from a crenulation cleavage to a local penetrative slaty cleavage at the micro-scale. Regionally S_2

strikes WNW to NNW with steep NE or SW dips but can be locally variable from later deformations. F_2 folds are tight and asymmetric and mostly plunge steeply. The D_3 deformation manifests itself as minor structures on S_2 and F_2, giving upright, open to tight, and generally symmetric folds and incipient crenulation cleavage. D_4 and D_5 occur in a few localities and are seen as close to coarsely spaced incipient cleavage or partings between which earlier planar structures are slightly to moderately crenulated. They are associated with eastward verging open to tight asymmetric (F_4) and gentle, upright, open (F_5) folds.

Metamorphism is low-grade greenschist containing chloritoid that may have succeeded D_5 deformation. Overprinting by contact metamorphism in the south appears to increase towards the Three Island granite.

The Triassic volcanic rocks have been separated into two types. One is a series of small flows and sills best seen along the west side of Whale Cove in the north, and the second is a 180 m (~500 foot) thick sill which extends over most of the island but is best seen along the western side. Zeolites such as heulandite, epistilbite, stilbite, and scolecite are common with natrolite, chabazite, apophyllite and laumonite more rare. Commonly the zeolites are coated with a green mineral which may be montmorillonite, chlorite or serpentine. Native copper may be found near the contacts of the various flows but is very rare. All are found in both rock types but are most accessible as well as abundant within the multiple flow-sill unit. The sill is characteristically columnar, jointed, as are the thicker smaller flows.

All major groups of rocks are separated by faults. The best exposed of these is near Red Head, where the Precambrian rocks are in fault contact with the Triassic red beds.

Detailed Descriptions

Stop 1: The Thoroughfare channel

The rocks at The Thoroughfare (Fig. 2) are very heterogeneous and difficult to describe because of compositional variations and diffuse boundaries. However, they are essentially black slate, grey phyllonite or phylonitic schist (meta-pelites), and quartzite cut by many generations of quartz veins and pods. To the north, past the lobster cages, these give way to weakly metamorphosed and deformed massive volcaniclastic rocks. All rocks are fine to medium grained. The primary layering is a transposed bedding-first foliation (S_0/S_1) that is a slaty to solution cleavage with 2 x 15 mm (average) microlithons. The structures are complex. The primary layering generally trends east-west, steep to south dipping, with later N-S trending asymmetric recumbent minor folds (F_2) and axial planar crenulation cleavage (S_2) (see Fig. 2).

The following structures have been recorded by P. Stringer (personal communication from an AGS Field Excursion to Grand Manan Island, July, 1991):

S_5: coarsely spaced crenulation cleavage, strike 100 , dips steeply North

S_4: crenulation cleavage, strike 015°, dip 14°-20° W

F_4: asymmetric open minor folds, trend 015°, verge eastward

S_3: crenulation cleavage, strike 120°, dip 80° S

F_3: minor folds, plunge SE at 35°

S_2: cleavage (much folded), variable strike E-W to NW-SE, dip 20°-80° S to SW

F_2: minor folds, locally distinguishable, plunge 40°-50° SE

S_0: (and subparallel S_1): bedding sheet-dip 20°-45° SSE.

Stop 2: Ingalls Head ferry terminal

The rocks at the ferry terminal are meta-volcanic mafic flows and sills, and volcaniclastic and tuffaceous metasedimentary rocks. They are all very highly deformed and chloritic in a manner similar to that seen at The Thoroughfare. For 300 m south along the coast from the jetty, primary layering between flows and sills and in volcaniclastic beds (subparallel S_0/S_1) dips northeast at 45°, followed south for 50 m by a structurally complex zone, in turn followed by ~50 m lacking outcrop. From 400 to 700 m south of the ferry jetty (as far as the point), the layering dips southeast with a nearly downdip lineation (L_1) defined by elongate chlorite amygdules in flows and dark fragments and elongate white clasts in the clastic units. These rocks have been refolded into late asymmetric open folds that trend N-S to NNE-SSW, verging to the east. In places there is coarsely-spaced crenulation cleavage associated with these folds. These late structures have been tentatively correlated with the F_4/S_4 structures at The Thoroughfare by Stringer.

Reference

Stringer, P. and Pajari, G.E. 1981. Deformation of pre-Triassic rocks of Grand Manan, New Brunswick in *Current Research, Part C, Geologial Survey of Canada,* Paper B1-1C, p. 9–15.

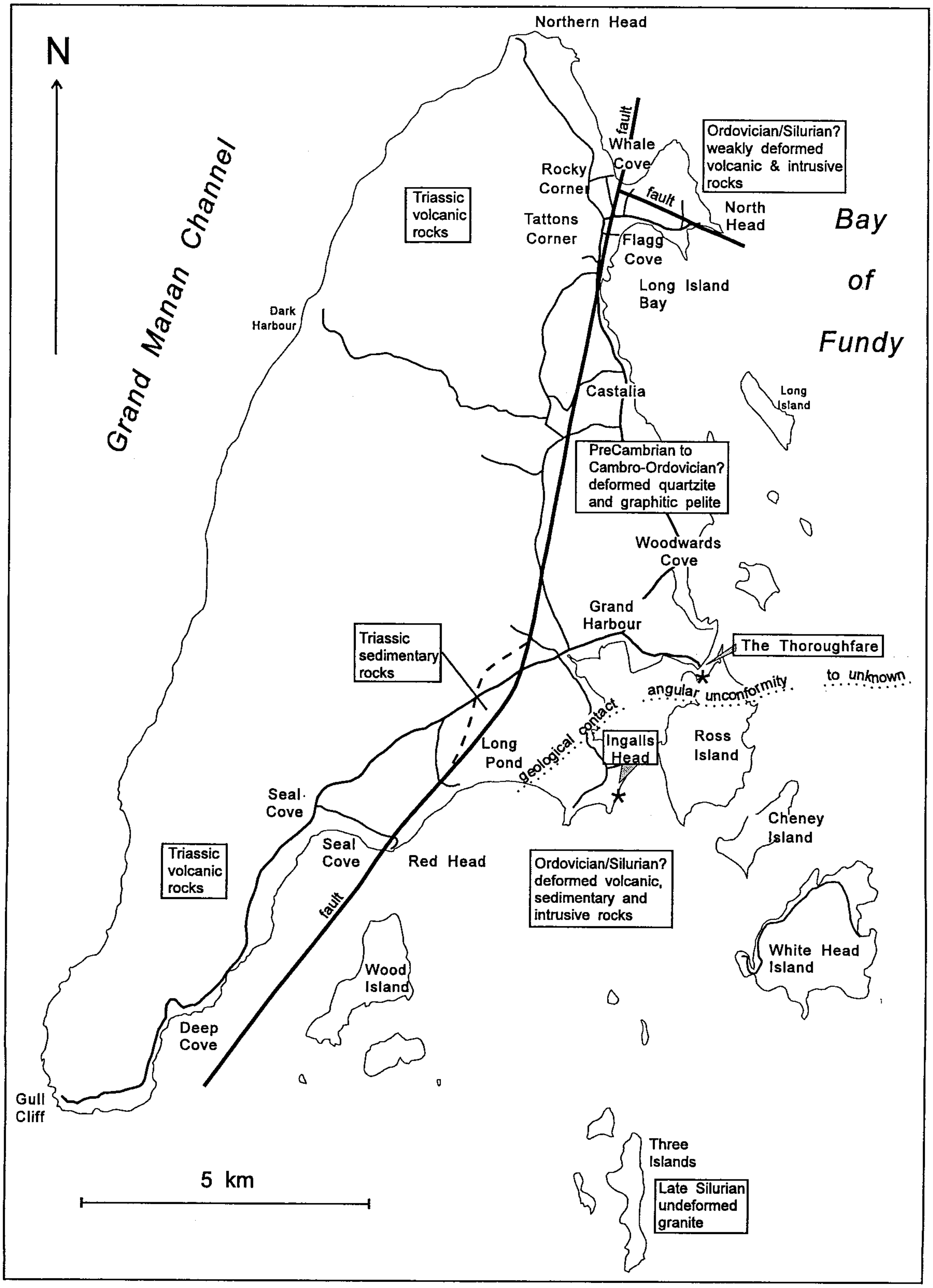


Figure 1: Location map of Grand Manan Island showing geological units and stop locations.

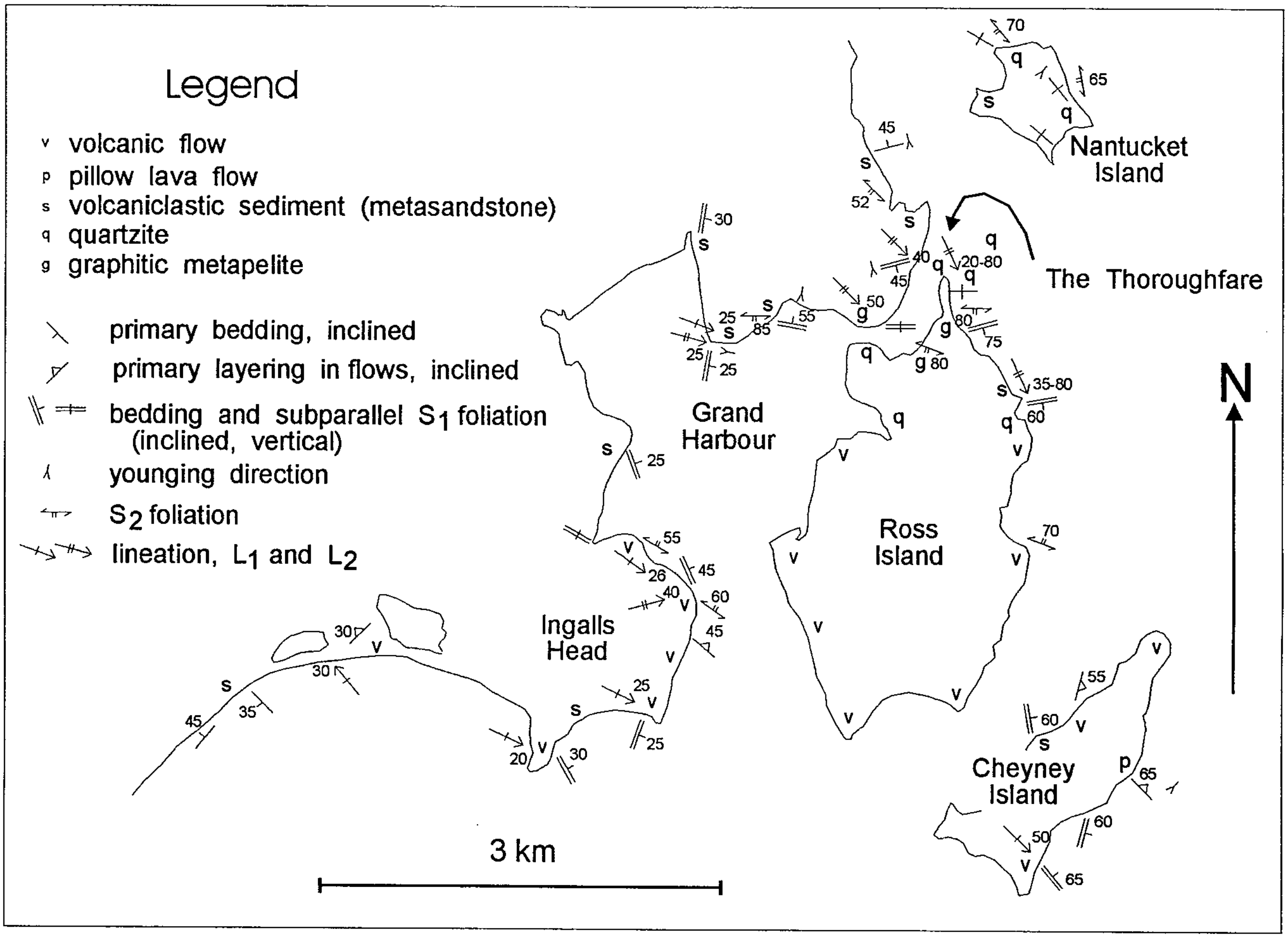


Figure 2: Geological map of the Grand Harbour area from the Thoroughfare to Ingalls Head showing distribution of lithology, bedding, and other primary layering, younging directions, secondary foliations (S2), and first and second tectonic lineations (L1 and L2). From Stringer and Pajari, 1981.

Fredericton

Bedrock Geology of the City of Fredericton including the Old City Quarry

Location

The two geological sites described in the field trip are located within the Fredericton city limits (Fig. 1). Stop 1 is located off Priestman Street opposite the hospital. It is just south of Regent Street in the Sainte-Anne Centre Communtaire, park near the tennis courts. The outcrop is just behind the courts towards the NB Tel tower. The property belongs to N.B. Telephone Company Ltd., P.O. Box 1430, St. John, NB, E2L 4K2.

Stop 2 is located on the north side of the Saint John River in Nashwaaksis, along highway 620 (Royal Road) about 2.5 km from the 105/620 intersection.At km 2.5 the road curves broadly to the north with outcrops on the north side. Just 10 metres before these outcrops is a gated dirt track. Park here or across the road out of the traffic. Be careful crossing the road and do not walk on the pavement.Take the dirt track to the Old City Quarry which is about 800 metres up and on the left fork. The first outcrops are within the road cut (10 m distance down 620) and are on Provincial property. The main Old City Quarry is registered in two names: Billings, Dawson R. and Billings, Audrey L., RR #7, Fredericton, NB, E3B 2X8.

Stop 1: Priestman St., Fredericton (view on St. John River)

The river at Fredericton has a channel about 1 km wide with bordering flats, in part floodplains, in part representing deltaic deposits of the Nashwaak and Nashwaaksis Rivers (Fig. 2). Valley walls rise relatively steeply from the flats on the south side, more gently northward, to an average height 80 metres above river level. The valley has a width of over 2 kilometres. The valley is a drowned one, the deepest part of the bedrock floor is as much as 70 m below river (and sea) level. Thus the real depth of the valley is nearly 150 metres. The present stream is relatively shallow, 3 to 10 metres, and runs over a filling of gravel, sand, silt, and clay, some of pre-glacial age, some glacial (Pleistocene), through mostly deposited in post-glacial times.

The bedrock is grey sandstone and minor conglomerate of the Pictou Group which is Pennsylvanian aged (Upper Carboniferous). These sedimentary rocks display irregular nearly flat bedding planes and are better exposed in the outcrop of the parking lot of the NB Tel Company below Stop 1 and on route across the river along the highway to Royal Road. Fossils of land plants were found in those rocks indicating an age of 250 million years old. The most abundant fossils represents stems of the rushlike plant Calamites, resembling modern horse-tails. Other fossils are imprints of the strap-like leaves of a primitive conifer, Cordites. Some large trunks of Cordites representing logs buried as the sediments accumulated were found in the Old Devon Quarry on Killarney Road. The wood has in part been converted to coal, and in part cells were infilled with silica and calcite, known as permineralization process. The minerals pyrite or marcasite may be seen in places. Brown and yellow staining in the rocks is limonite. The fossils of land plants and irregularity of bedding and variation in grain size, indicate that the beds are of non-marine origin, laid down in channels and floodplains of meandering streams and related lakes and swamps.

More recently a variety of erosional landforms were produced as a glacier attempted to streamline its bed, particularly on the surface of the Carboniferous sandstone and conglomerate. Two types of features are common: positive and negative (Fig. 3). All these features can be seen on the glaciated surface off Preistman Street with striations being the most common and all the others more rare. Sometimes it is better to view these features after a rain while the rock is still wet.

Stop 2: Old City Quarry, Royal Road

Beds of grey sandstone and conglomerate are exposed in the road cut of the Royal Road, as it descends to the floodplain of the Nashwaaksis River from the junction with the high level entry to the quarry. Conglomerate is more prominent here. Pebbles of white quartz, well-rounded are predominant. The beds here are inclined in contrast to the flat attitude of the road bed. The angle between the bedding planes and the horizontal is known as the dip of the beds, here around 45° to the east. Inclined beds like these are intersected by the ground surface and this trend across country may be traced and its direction measured. The trend across a horizontal surface is known as the strike of the beds, here around N 030 W. The beds are inclined because of disturbance since their deposition; the disturbance is associated with slipping along a nearby fault (Fig. 2).

The beds when laid down were nearly flat-lying and younger beds would have been deposited atop older ones. As the beds are now tilted towards the east successively younger beds and formations are met as one walks westward and north-westward along the quarry road. The grey sandstones are exposed at intervals along the road leading up to the quarry.

Overburden consisting of glacial moraine covers the bedrock of sandstone and conglomerate between outcrops. Some large boulders of granite are to be seen. They are called erratics because they do not match the underlying bedrock. The nearest granite is about 16 kilometres to the northwest and is the most likely source for the erratics since we know from other evidence that the Pleistocene ice-sheets advanced from that direction.

Along the dirt track near to Royal Road, beds of grey sandstone are in contact with beds of conglomerate, a good place to measure the strike and dip. Further along we find a change in the bedrock, a mottled green shale underlies the grey sandstone and conglomerate formation. Note the smooth feel of the shale, its thin beds or laminae, and its tendency to break with a somewhat conchoidal fracture along the laminae.

A gully crosses the road a few feet beyond the shale exposure. The gully, occupied by running water after heavy rain or melting snow, is a drainage channel for the adjacent ground, and has been eroded by the running water. Great streams, such as the St. John River and nearby Nashwaaksis start out as gullies and gradually deepen and widen with time and as more and more run-off is diverted into them. Note the V-shaped cross-section and irregular gradient of the gulley characteristic of youthful streams.

A few metres up the gulley is an exposure of a third formation. The weathered surface is brownish but freshly-broken rock is dark green. Its texture is fine-grained. Vesicles are present, some infilled with calcite and a green mineral chlorite. The filled vesicles are known as amygdules. Vesicles indicate a rock of volcanic origin and mark escaping gas entrapped in quickly cooled lava. Colour and density of the rock suggest a basic composition. The rock is basalt, sometimes called trap rock.

The basalt is poorly exposed to the base of the quarry where it is underlain by a pebbly red sandstone. The sandstone in turn overlies another lower basalt which we will find to be much thicker than the upper one, over 20 metres in part, as compared to about 7 metres for the upper flow and about 7 metres for the red and green shale formation (Fig. 2).

Farther along we have the deeper, middle part of the lower basalt exposed in the quarry. Vesicles are absent here and the rock is medium-grained due to slower cooling. Nearly vertical fractures or joints cross the basalt dividing it into indistinct, crude columns (columnar jointing).

A polished vertical surface forms part of the wall of the quarry. Examination shows it to have been a prominent fracture. Horizontal lines along it, known as slickensides, indicate that one wall of the fracture has been shifted horizontally against the other wall, a structure known as a fault. The material lining the fault is in part calcite, and where shiny, a metallic grey mineral which gives a red streak on scratching. The mineral is specularite, the metallic, grey, variety of hematite. Some red hematite is also present along this and other fractures.

We can follow the basalt to where the bottom is exposed in the quarry where we can see the red sandstone onto which it was placed in the Mississippian and/or Pennsylvanian times. A few vesicles are present. Cooling was quick enough here to trap some escaping gas bubbles and to produce a fine-grain, although not to the same extent as at the top of the flow. The underlying sandstone shows hardening, or baking, due to the heat of the lava for a few cm below the contact.

On top of the basalt, we can see again the vesicular top and also the overlying pebbly red sandstone. The uppermost part of the basalt is strongly broken and weathered. This gives us evidence that the basalt was formed as a flow. Above the red sandstone and conglomerate is exposed the other smaller basalt flow. Here we also see scratches on the lava surface made by rock fragments carried by the ice sheets of the Pleistocene epoch. These striations tell us the direction (SE) of the ice movement.

From the top of the quarry we get a good view of the deep valley of the Nashwaaksis. Note the relatively small present stream channel on the west side of the valley. The stream once started as a gully over our heads and gradually eroded downward and meandered back and forth to carve out its present large mature valley; the bedrock bottom is, as for the St. John River valley nearly 70 metres below the present stream valley, and like the St. John was depressed below sea level during the glacial period, and infilled with glacial and post-glacial deposits.

Flats bordering the channel represent material (alluvium) laid down during floods of relatively recent times and hence are called floodplains. Traces of former locations of the channel may be discerned on the flats.

Contibutors

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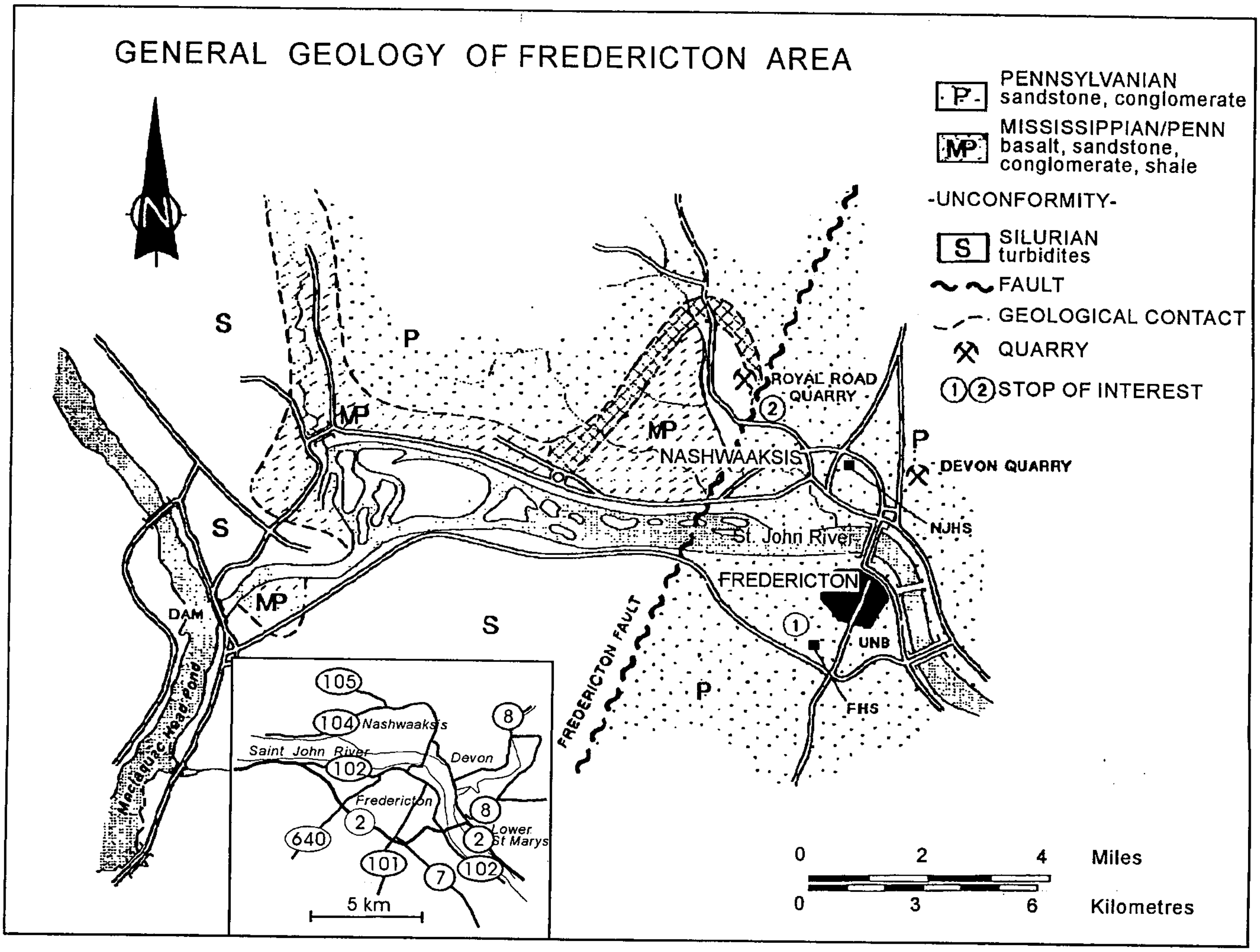

Figure 1: General Geology Map of the Fredericton area showing the locations of the stops.

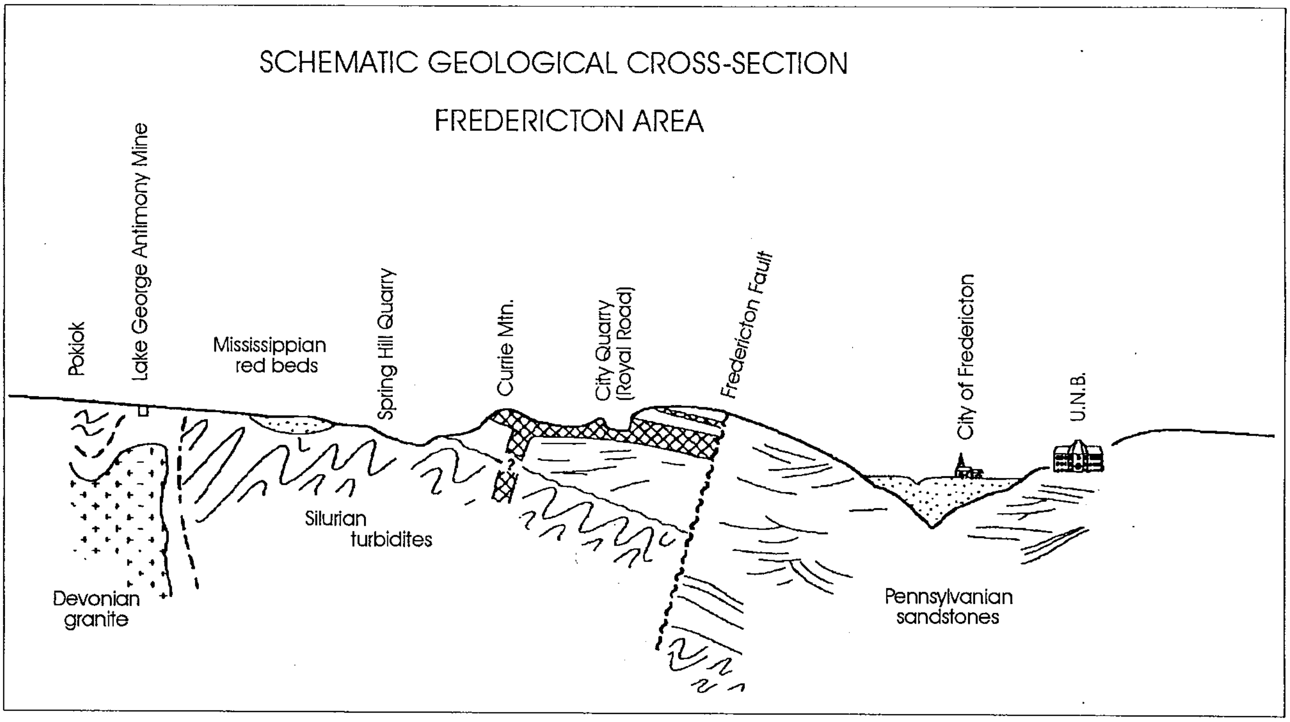

Figure 2: Schematic geological cross-section oriented NNW starting at UNB to past stop 2 on Figure 1.

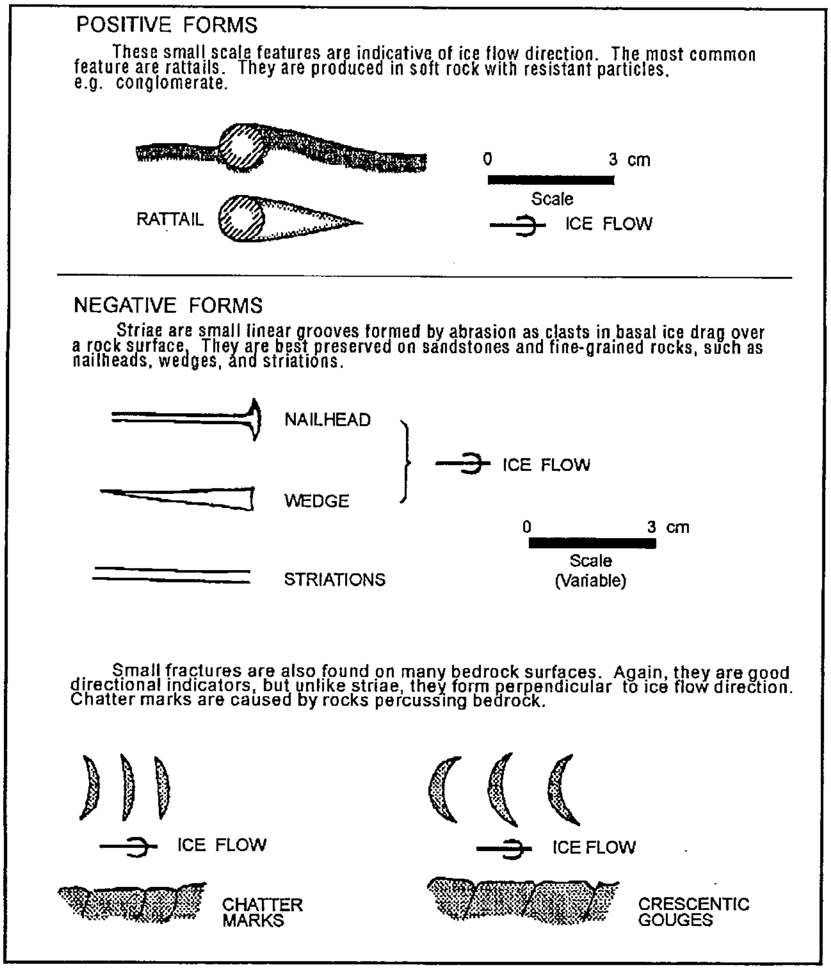

Figure 3: Positive and negative erosional landforms resulting from glaciation.

The Bathurst Mining Camp

Nepisiguit River Falls and Austin Brook Iron Mine

Purpose

These two stops provide an overview of the stratigraphic and structural relationships of some of the rock units in the Bathurst Camp, namely those associated with the Brunswick No. 6 and No. 12 mines although the mines themselves will not be visited.

Location

Take Route 430 south from Bathurst or from 360 west off Highway 8 (Fig. 1). Go past the road to Brunswick No. 12 for about 0.7 km and keep on straight at the turn (the turn goes to Heath Steele Mines). Take the next left about 1 km and continue past village to a T-junction about 3 km along. Drive ahead onto the dirt road to the left of the electric generating building and park off the road for the first stop (Fig. 2). The roadbed that parallels the Nepisiguit River is all that remains of the Northern New Brunswick and Seaboard Railway line that went from Newcastle to the old Austin Brook Iron Mine in the early 1900s.

The Austin Brook iron mine (Figs. 1 and 3), the second stop, is to west of the T-junction about 4 km down the dirt road past the dammed lake. Go past the pump house for the mine and across a small brook (Austin Brook). Park at the first dirt road to the left (a distance of about 150 m), and walk about 150 m up the hill on the dirt track to the mine. Go to the southwest end where there are glaciated outcrops.

The road continues to Brunswick No. 6 (1966-1983; named after the exploration project) which is less than 900 m away but there is no access to the mine so you must turn around to go back.

Introduction

The Tetagouche Group is characterised by an anomalous abundance of Zn-Pb-Cu massive-sulphide deposits. At present (1997), base metals are mined from the Brunswick No. 12, CNE and Heath Steele deposits, whereas Brunswick No. 6, Wedge and Caribou are past producers. Gold and silver were extracted from gossan overlying the Murray Brook, Caribou and Heath Steele deposits, and in the early part of this century the iron formation in the hanging wall of the Austin Brook massive-sulphide deposit was mined for iron.

Stratigraphy

The rocks in the Bathurst Camp, part of the northern Miramichi Highlands, have been separated into three groups, namely Miramichi, Tetagouche, and Fournier. The Tetagouche Group conformably to disconformably overlies the Miramichi Group and is structurally overlain by the Fournier Group. The Tetagouche Group hosts most of the massive-sulphide deposits.

The Middle to Late Ordovician Tetagouche Group comprises five formations: Patrick Brook (PB), Vallee Lourdes (VL), Nepisiguit Falls (NF), Flat Landing Brook (FLB), and Boucher Brook (BB) in ascending stratigraphic order, more or less. The PB and VL formations are sedimentary units that occur locally at the base of the group. The conformably overlying NF and younger FLB formations form a thick felsic volcanic pile and constitute the most voluminous part of the group. The felsic volcanic pile ranges in composition from dacite to rhyolite. Sedimentary and mafic volcanic rocks of the BB Formation conformably overlie the FLB Formation in the type area. A sixth unit of mafic volcanic and minor sedimentary rocks, called the Canoe Landing Lake (CLL) Formation, is in thrust contact with the type-BB Formation. However the CLL appears to be coeval with, and identical to the Boucher Brook. Therefore, the CLL is considered to be part of the Tetagouche Group.

Structure

The structural geometry of the Bathurst Camp reflects an interference pattern produced by polyphase deformation, something that was recognized by early workers. Detailed structural analysis and overprinting relationships has led modern workers to recognise five groups of folds, of which the first two sets are responsible for most of the complex geometry.

The earliest deformational event (D_1 - Late Ordovician to Early Silurian) is represented by steeply inclined to recumbent, non-cylindrical folds (F_1) with an axial-planar, layer-parallel transposition foliation (S_1), and generally a stretching lineation (L_1). During the second phase of deformation (D_2 - Late Silurian) S_1 was reoriented into a near-vertical attitude by tight to isoclinal F_2 folds. The S_2 cleavage is moderately to well developed and generally steeply-dipping. Along the limbs of the F_2 folds, S_1 and S_2 are sub-parallel and may form a composite S_1/S_2 cleavage (S_{MAIN}). The post-D_2 structures include a recumbent folding phase followed by two sets of folds that range in scale from millimetres to kilometres, and produce dome and basin structures. Regardless the S_1 and S_2 cleavages are the dominant fabric elements seen throughout the area.

Tectonic Setting

The tectonic setting of the Bathurst Camp is that of a back-arc basin as depicted in Fig. 4. The volcanic rocks of the Tetagouche Group are believed to have been laid down as basin-margin deposits on rifting continental crust, and the Fournier Group rocks represent oceanic crust that formed during spreading of the basin. From the Late Ordovician to Early Silurian this back-arc basin started to close by northwest-directed subduction and the Tetagouche Group rocks are thought to have been underplated to the oceanic part (Fournier Group) of the accretionary wedge when the leading edge of the continental margin descended into the subduction zone. Closure of this basin culminated with the obduction of trench-blueschist onto the former margin of the basin. Within this tectonic scenario, D_1 and M_1 are seen as the products of major thrust faulting and metamorphism (high pressure and low temperature) in the subduction-related closure of the oceanic basin. Post-D_1 ductile deformation resulted from the subsequent oblique, more or less continuous collision between Laurentia and Avonlonia, which ended in the Lower Devonian.

Massive-sulphide Deposits

Massive sulphide deposits occur in several stratigraphic positions within the Tetagouche Group. Many are closely associated with fine-grained sedimentary and/or felsic volcaniclastic rocks of the Nepisiguit Falls Formation; others are in the lower sedimentary part of the Boucher Brook Formation and a few are within the Flat Landing Brook Formation. The deposits, which range in size from small showings to supergiants such as Brunswick No. 12, comprise concordant lenticular, massive to disseminated bodies that mainly consist of pyrite, sphalerite, galena, chalcopyrite, magnesite and, in places, pyrrhotite. Other sulphides (particularly arsenopyrite), sulphosalts, and iron oxides occur in minor amounts.

Most of the major deposits are hosted by fine-grained sedimentary and/or volcaniclastic that are generally chloritic and comprise a mixture of exhalative and epiclastic components, the proportions of which vary along strike. The areal extents of these fine-grained rock units are commonly much greater than the lenticular to sheet-like sulphide deposits enclosed within them.

In this area the deposits occur in the Brunswick Horizon at, or close to, the contact between the Nepisiguit Falls and Flat Landing Brook formations. These Brunswick-type deposits generally are associated with Algoma-type iron formation and include the Brunswick No. 12, Brunswick No. 6, Austin Brook iron mine, Flat Landing Brook and Key Anacon deposits.

The preferred depositional model for the large massive sulphide deposits of the Bathurst Camp is a density-stratified brine pool that formed in relatively deep water (500-1000 m). The metalliferous fluids were largely generated by stratabound flow of seawater through the devitrifying NF volcaniclastic pile.

Outcrop Descriptions

Nepisiguit Falls dam

The outcrops in the area of the dam (Fig. 2) constitute the type section of the Nepisiguit Falls Formation. Granular texture is apparent in this outcrop of quartz-feldspar-augen schist (QFAS), which predominantly consists of juvenile volcaniclastic material with a few accidental lithic fragments. This type of QFAS constitutes part of the proximal volcaniclastic facies and by comparing textures in the various outcrops one can detect variations in the grain size and abundance of quartz and K-feldspar phenoclasts. In the water-polished outcrops at the foot of the dam, thick crudely-graded beds can be seen. These rocks are interpreted as cold debris flows rather than hot pyroclastic deposits.

About 100 m downriver, along the railway bed are several road cuts on the north side. In the first part of the rock-cut on the right, a thin layer of cherty hyalotuff caps finings upward QFAS. This hyalotuff represents fine-grained glass particles that separated from the crystal-rich debris flow during its emplacement, and then settled from the water column before deposition of the next flow. These hyalotuff beds are rare in the proximal facies of the formation but predominate in the distal facies.

The "Brunswick Horizon" can be seen in the ditch along the road past the dam. In this poorly exposed outcrop that runs for about 100 m, massive rhyolite of the Flat Landing Brook Formation is in contact with the chloritic iron formation of the Nepisiguit Falls Formation. Note the contrast in cleavage development in these two rock types. Some of the chloritic rocks are magnetic and manganiferous reflecting their original, chemical-sedimentary character, whereas others exhibit remnant volcaniclastic textures indicating that they are the product of hydrothermal alteration.

Austin Brook Iron Mine

The deposit (Fig. 3) was discovered in 1902 and in 1910 was actively mined by the Canada Iron Corporation with the first ore shipped to Newcastle by way of the railway. It was closed in 1913 and opened again in late 1942 when, for about one year, 130,000 tons were mined and shipped to Sydney, Nova Scotia.

The ore has a general black colour, tinged greyish from the presence of minute, pale grey minerals (namely quartz and feldspar). The ore body is fine to lenticular, layered in varying degrees from microscopic to very broadly developed, the layered appearance is due to alternations of rather minute, band-like streaks of magnetite, quartz, and feldspar, with varying, much smaller amounts of calcite, chlorite, biotite, sericite, and hornblende. Towards the footwall, which is itself heavily impregnated with sulphide, the ore contains considerable pyrite, while, more particularly towards the centre of the mass, considerable quartz is present in veins and stringers. However the bulk of the ore consists of quartz and magnetite, both in a finely granular state.

Pyrite occurs in fine grains and in minute parallel lines and streaks, usually very short, but in some cases several centimetres or so in length. Near the footwall, the pyrite is less uniformly distributed, and tends to occur in very narrow, distinct, vein-like masses of comparatively large grains showing distinct crystal faces. Boundaries between sulphides and iron oxides are irregular but sharp with the sulphides being almost completely free of iron oxide and the magnetite pyrite. In some places ore and sulphide are regularly interbanded, the ore dense, the sulphide comparatively coarse grained.

The ore body has the form of an abruptly terminating bed with a fairly constant thickness. The walls are sharply defined, and dips westward at a moderate to steep angle. The rocks, in general, have a prominent slaty cleavage parallel to layering dipping about 45°-90° to the west. Only one cleavage is seen and it is probably a composite layering S_1/S_2. Small (centimetre to decimetre scale) intrafolial and refolded folds are seen in many of these layers. The ore body is cut by numerous narrow faults with offsets of a few cm or less.

The footwall is a highly weathered sericitic quartz augen schist, very heavily charged with pyrite, and has a very sharp contact with the ore body. It ranges in width from a few centimetres to six metres. The hanging wall is rhyolite of the Flat Landing Brook Formation and it too has a very sharp contact.

Source

McCutcheon, S.R., Langton, J.P., van Staal, C.R. and Lentz, D.R. 1993. Stratigraphy, Tectonic Setting and Massive-Sulphide Deposits of the Bathurst Mining Camp, Northern New Brunswick, p 1- 28. *in Guidebook to the Metallogeny of the Bathurst Camp*, S.R. McCutcheon and D.R. Lentz, eds., Trip #4 of Bathurst '93: 3rd Annual Field Conference, Geological Society of CIM.

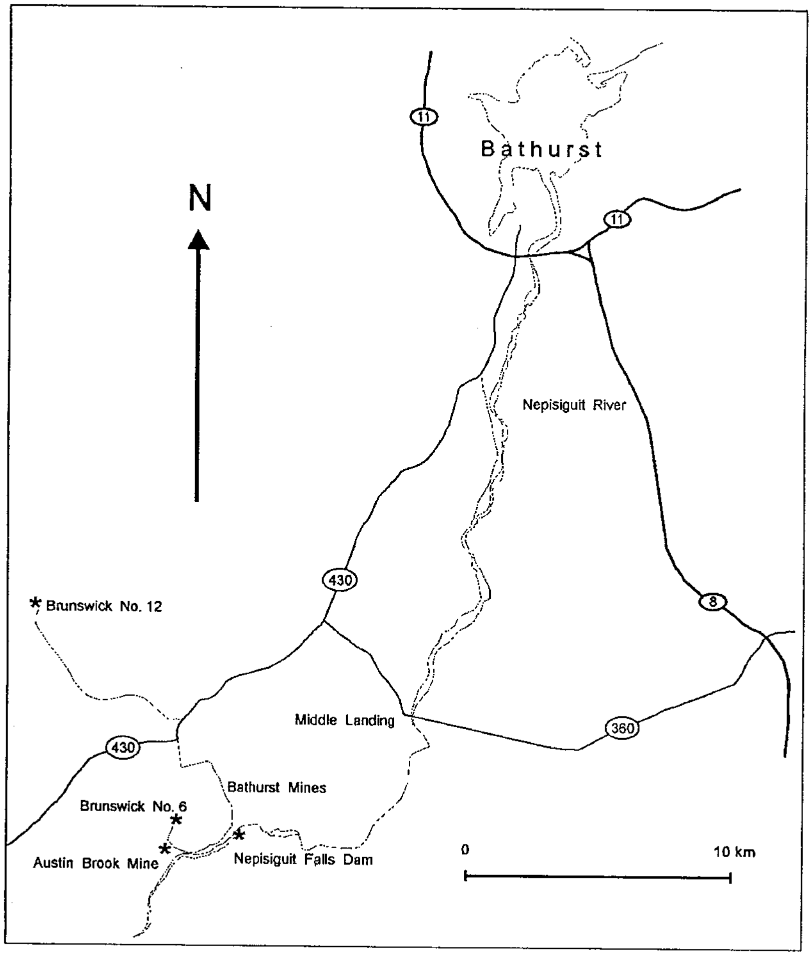

Figure 1: Location map for Bathurst Mining Camp

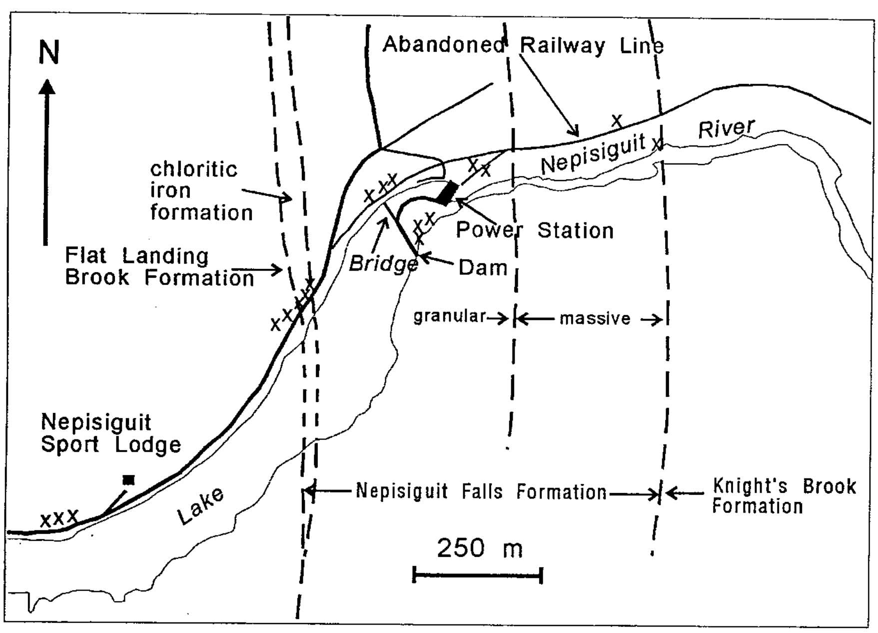

Figure 2: Simplified geological map of the Nepisiguit Falls area. Modified from McCutcheon et al., 1993.

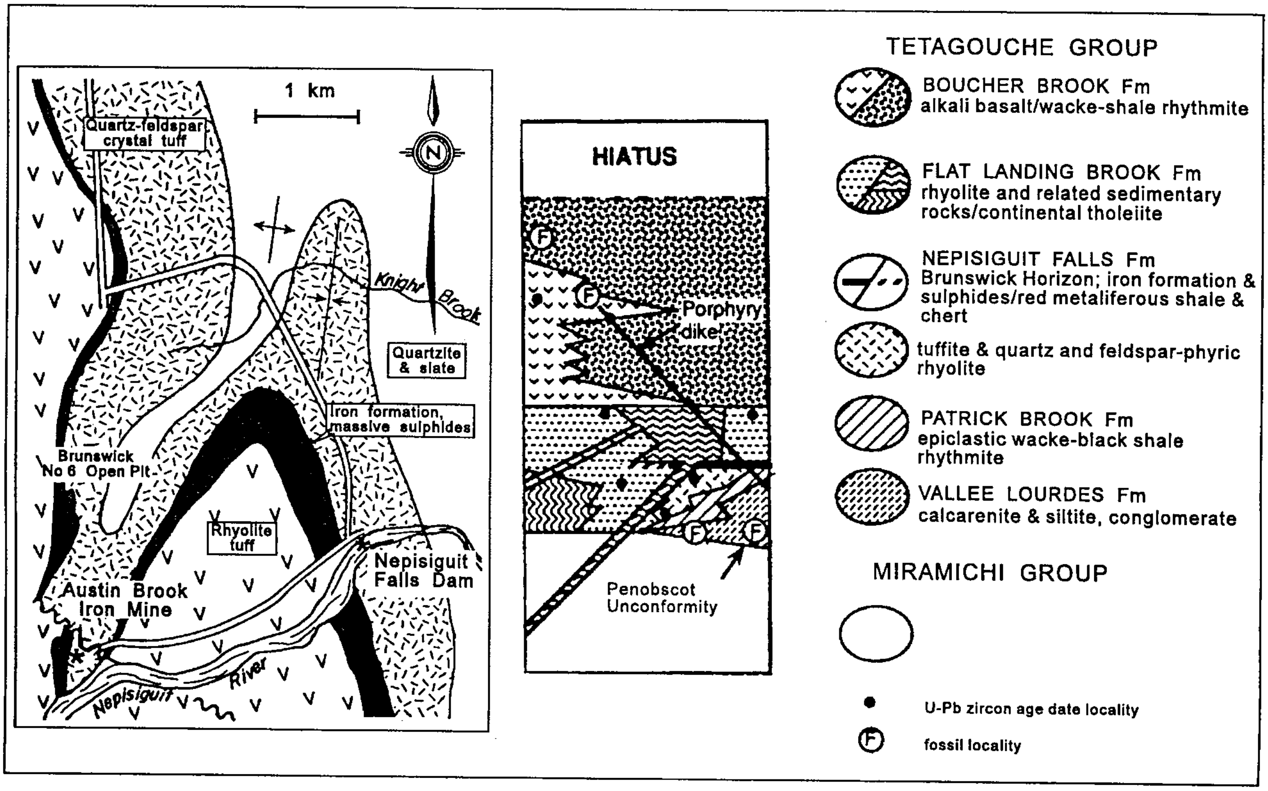

Figure 3: Simplified geology map and general stratigraphic column for the Nepisiguit Falls dam and Austin Brook mine areas. Modified from Fyffe and Noble, 1985, and Goodfellow and Peter, 1996.

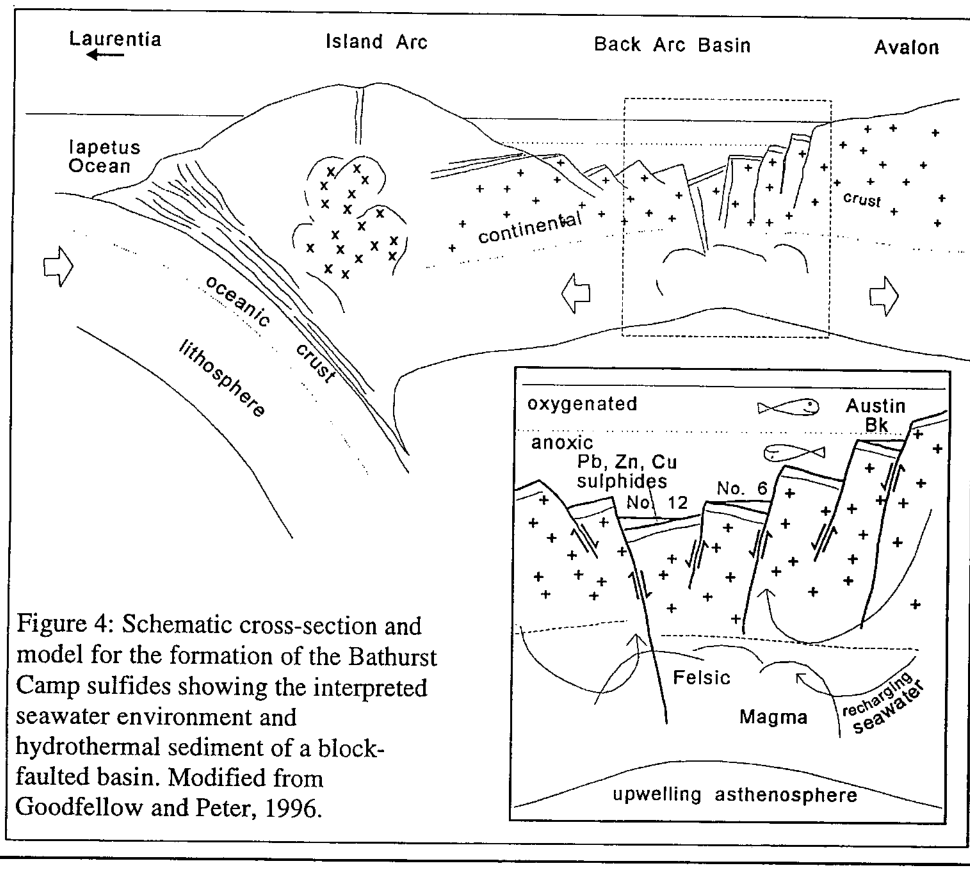

Figure 4: Schematic cross-section and model for the formation of the Bathurst Camp sulfides showing the interpreted seawater environment and hydrothermal sediment of a block-faulted basin. Modified from Goodfellow and Peter, 1996.

Pabineau Falls

Pabineau Falls Granite

Purpose

One of the best exposures of granite in the area.

Location

Take Route 430 south out of Bathurst and about 5 km past the Highway 11 overpass take the left turn towards the Pabineau River Indian Reserve. The paved road continues straight, changes into gravel until it comes to the falls on the Pabineau River, at about 2.8 km from the turn. Beyond the falls the road is a dirt four-wheel drive road (Fig. 1).

This location is on native land so respect of private property is a necessity. It is also next to dangerous rapids and falls, so extreme caution is advised. Do not go close to the water and be careful of where you walk.

Description

The Pabineau Falls granite is a red hornblende-biotite monzogranite. It is megacrystic with 2-10 mm anhedral megacrsyts of K-feldspar in a finer grained groundmass of 2-5 mm K-feldspar, quartz and minor plagioclase and 1 -3 mm hornblende and biotite. Randomly scattered throughout the outcrop are thin layers of varying composition or schlieren ranging from monzogranite to diorite. Mineral variables are K-feldspar, quartz, plagioclase and hornblende, biotite appears to be constant. Associated with or without these layers are thin to wide (1 mm to 10 cm scale and up to 1 m wide) zones of foliation.

There are a few round, small (10 cm diameter), mafic inclusions with igneous textures and the granite is cut by thin to thick (1–3 cm average but up to > 1 m) aplite veins and dikes. These are generally oriented east-west and vertical and extend the whole length of the outcrop. One dike shows multiple intrusive phases or layering of aplite and fine grained granite. Contacts are sharp and straight with grain sizes uniform throughout. Pods of pegmatite with or without minor aplite containing equal amounts of quartz and K-feldspar, 10-30 cm in diameter, also occur. Contacts are diffuse or gradational zones less than 1 cm wide.

The whole outcrop is cut by meter-scale vertical and closer spaced horizontal exfoliation joints that are selectively weathered.

The granite has been dated using zircons as 396 ± 4 Ma.

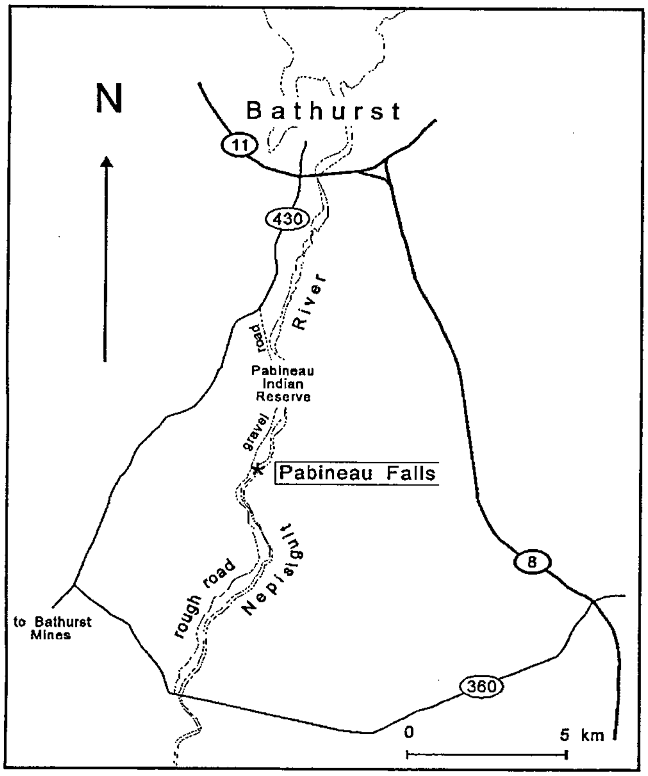

Figure 1: Location map for Pabineau Falls

Bathurst

Tetagouche Group Pillowed Basalt

Purpose

To view a superb outcrop of pillowed basalt.

Location

From Bathurst take Highway 11 north. About 2.7 km north of the Route 315 overpass (just past the Tetagouche River) there is a gated dirt road to the west (Fig. 1). This road goes into the now disused portion of the St. Isadore Quarry. Park outside the gate and walk in to the first large glaciated outcrop to the north (approximately 200 m).

Introduction

The Tetagouche Group in northern New Brunswick is a sequence of early Ordovician aged metasediments followed by intercalated felsic volcanics and metasediments including iron formation and minor limestone and capped by Middle Ordovician pillowed basalts interbedded with slate, greywacke, and limestone. Chemical analysis of the Tetagouche volcanics reveal a bimodal distribution with a conspicuous absence of andesitic compositions. The basaltic rocks are spilitic with titanium contents indicating both alkaline and subalkaline classifications. The subalkaline basalts are tholeiitic with ocean floor affinities and the alkaline basalts are intraplate. The felsic volcanics range in composition from dacite to rhyolite. The broadly circular distribution of the felsic volcanic rocks in the Bathurst area suggests the presence of a large caldera complex and their association with volcanics and granites of both Ordovician and Devonian age implies that the area is a rifted fragment of continental crust.

The most prevalent structures recognised within the Tetagouche Group are tight to isoclinal mesoscopic folds possessing a penetrative axial planar cleavage. Large folds recognized by regional mapping of contacts between lithologies are probably equivalents to these outcrop scale mesofolds. An earlier transposition of layering with associated intrafolial folds are occasionally observed. Metamorphic grade is prehnite-pumpellyite facies in the Bathurst region and increases to the southwest.

The Quarry

The rocks of the quarry contain large, well-preserved basaltic pillows and massive flows. The pillow shapes vary from "loaf-like" to "mattress-like", give top indicators to the north, and contain variable amounts of amygdules (up to 20 per cent). Rinds are 2–3 cm thick in most places and there is a crude east-west layering of the pillows. Locally the pillows have red chert between them. The basalt is variably porphyritic with zoned subhedral plagioclase phenocrysts up to 3.5 mm in length. The pillows show zonation of grain sizes and vesiculation from the margins inward and some pillows have pipe vesicles near their tops and/or bottoms. The rocks are locally cut by epidote veins and have been glacially planed with striae oriented 215-035°.

The age of the basaltic rocks can be determined from graptolites found in underlying black slate just upstream from the railway bridge over the Tetagouche River (see Fig. 1). The locality can be reached by walking through the gravel pit on the east side of the bypass and following a foot path above the wall of the pit down to an old logging road. Follow the logging road downstream to where it meets the river and continue along the river's edge to the outcrop of black slate within sight of the railway bridge. The rocks are of mid-Caradocian age.

Source

Fyffe, L.R. and Noble, J.P.A., 1985. Stratigraphy and Structure of the Ordovician, Silurian, and Devonian of Northern New Brunswick. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Fredericton, '85, 56p.

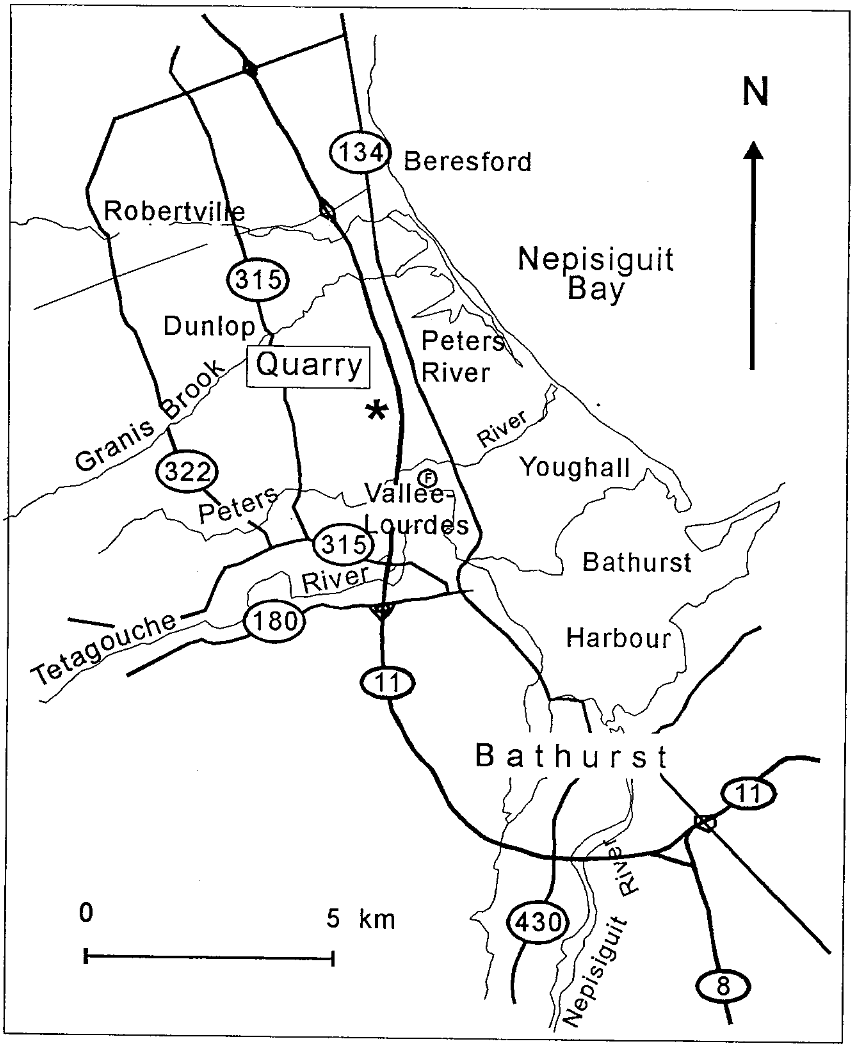

Figure 1: Location map for Tetagouche Group pillowed basalt as well as the graptolite fossil location (F).

Chapel Point

Red Boulder Conglomerate of the South Charlo Formation and the Bryant Point Volcanics

Purpose

To view the red boulder conglomerate of the South Charlo (formerly New Mills) Formation and Bryant Point volcanics.

Location

From Bathurst take Highway 11 north to Belledune and exit the highway to route 134 north (Fig. 1). Exit Curry Road off 134 and go south towards the NB Power Plant. Stop before the fence around the plant (north side) and take the small dirt track to the beach. Go south on the beach to the rocks at the point (Fig. 2).

Introduction

Beneath the Carboniferous cover of New Brunswick the older rocks are divided into a number of tectonostratigraphic zones with distinctive stratigraphy and structural styles. In northeastern New Brunswick these zones are the Miramichi Massif and the unconformably overlying (?) to faulted Matapedia Basin. They outcrop on the coast from Bathurst northwards and display the older mid-Ordovician Taconic and mid-Devonian Acadian Orogenies respectively. The Chapel Point rocks are part of the Silurian aged Chaleurs Group which are composed of a shallowing-upward sequence of Lower to Middle Silurian siltstone and sandstone locally overlain by nodular limestone and Upper Silurian mafic volcanics. These are followed conformably by calcareous siltstone, shale, and limestone interbedded with mafic volcanic flows, palagonite tuff and minor felsic volcanic rocks of the Lower Devonian aged Dalhousie Group.

The Siluro-Devonian rocks possess a single generation of open to close folds with shallow plunges to the northeast or southwest with some local steepening. A steep slaty cleavage, well developed in all but conglomeratic and volcanic rocks, can be seen to be markedly oblique to axial surfaces in exposures along the shoreline. Stringer (1975) believes this to be the result of a very early initiation of irregular buckling of strata prior to the onset of cleavage development during northwest-southeast directed compression. Most rocks have undergone zeolite facies burial metamorphism.

Chapel Point

At this stop (Fig. 2) it is possible to see a thin section of Bryant Point volcanics and the South Charlo (formerly New Mills) conglomerates and sandstones. The section is complicated by a steep northeast-southwest fault, but most of the section can be examined west of the fault. The Bryant Point volcanics are dominantly amygdaloidal basalts which have a very distinctive porphyritic texture. Variable phenocryst content ranges from > 50% to < 10%; the phenocrysts are skeletal, subhedral to anhedral, and average 6mm in size with some as large as 15 mm. Many areas show the phenocrsyts to be agglomerated and many have grown together in a radial pattern. Zonation of amygdules within the basalt are reminiscent of pahoehoe lobes. The top of the basalt is weathered and eventually grades into sandstone and siltstone which is cut by the unconformably overlying Carboniferous Bathurst or Boneventure Formation.

The South Charlo conglomerate consists of about 16 m of coarse boulder conglomerates passing up into red sandstones. The clasts, averaging 20 cm diameter and ranging up to 80 cm, are mainly basalt plus fossiliferous limestone from the Bryant Point and lower La Veille Formations with a few chert and other "exotics". Clasts are mainly matrix-supported and rounded. Crude bedding seen in a few areas >40 cm thick, is defined by grain size variation and imbrication of the prolate clasts which indicates currents from the west. Included are some finer grained red calcareous sandstone beds 20 - 40 cm thick, with large cross-beds and megaripples and channel cut bottoms. These show current directions from the northeast. The clast lithologies indicate considerable local uplift and erosion prior to South Charlo deposition and the top indicators show the beds to be overturned to the south.

The volcanics have numerous zones and veins of epidote mineralization and the conglomerate has an incipient fracture to slaty cleavage dipping moderately to the southeast. Interestingly some of the basalt boulders in the conglomerate show the cleavage better than the rest of the unit while the Bryant Point volcanics show no cleavage development.

The South Charlo conglomerate can be dated approximately as Pridolian to earliest Devonian, and the Bryant Point volcanics as slightly earlier or synchronous. The evidence suggests that this was a time of rapid uplift and erosion.

The unconformably overlying deep red Carboniferous rocks are exposed to the south towards the Power Plant. They are nearly flat lying, crudely bedded pebble conglomerates and interbedded sandstones.

Source

Fyffe, L.R. and Noble, J.P.A., 1985. Stratigraphy and Structure of the Ordovician, Silurian, and Devonian of Northern New Brunswick. Geological Association of Canada Mineralogical Association of Canada, Joint Annual Meeting, Fredericton '85, Excursion 4, 56p.

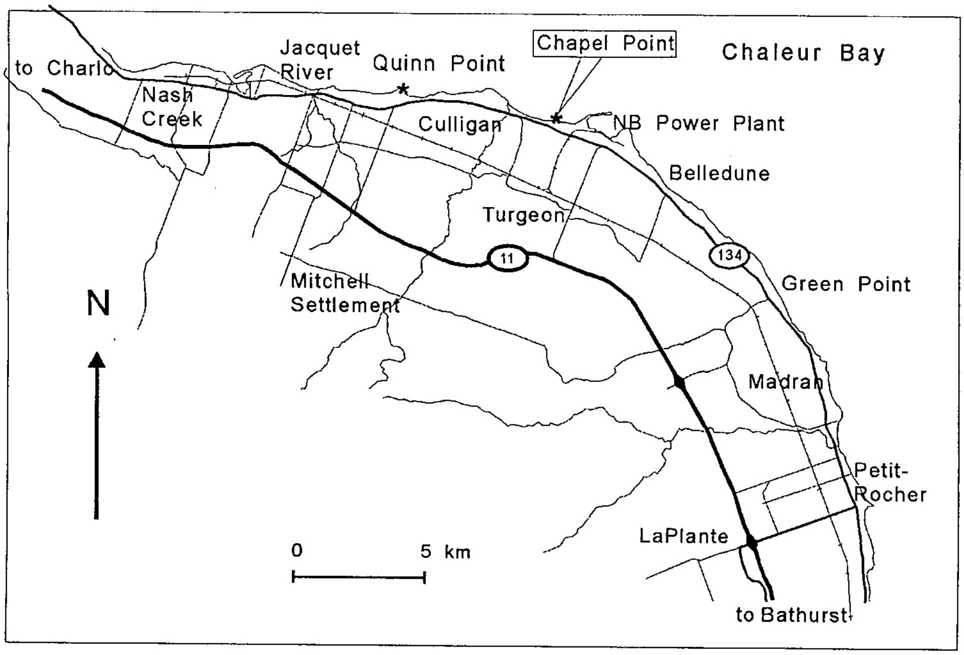

Figure 1: Location map of Chapel Point

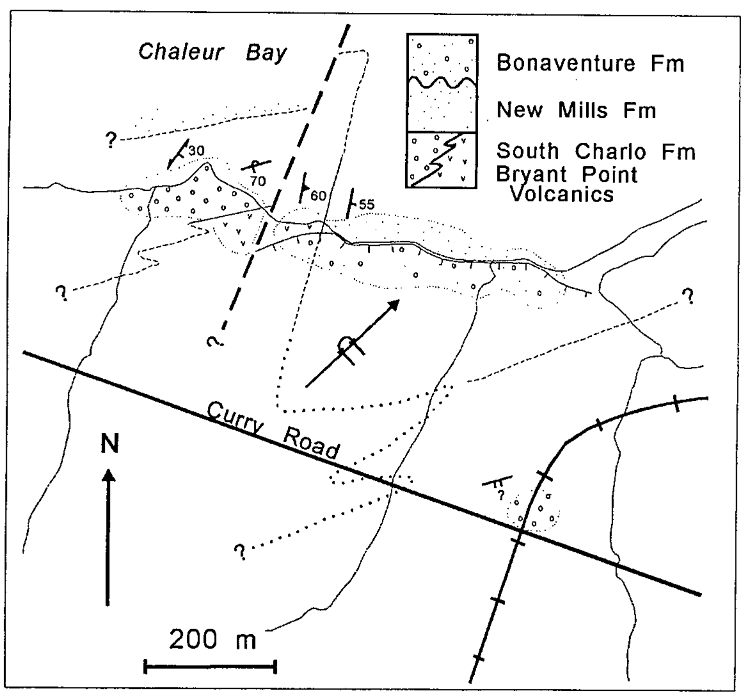

Figure 2: Detailed geology map of Chapel Point area.

Quinn Point

Sedimentary Rocks and Stromatoporoids of the Silurian aged La Vielle Formation

Purpose

The section exposed at Quinn Point show a Llandoverian-aged transgression from continentally derived sediments to deep shelf carbonates followed by a Wenlockian regression back to intertidal and/or supratidal algal stromatolites. Also some of the best exposures of the Stomatoporoid biofacies of the La Vieille Formation are exposed at Quinn Point itself.

Location

From Bathurst take Highway 11 to Belledune and then go north on Route 134 (Fig. 1). About 3 km north of Belledune at the village of Armstrong Brook take Smith Road towards the Chaleur Bay shore. The road ends before the beach, park and take the dirt road down to the beach. Quinn Point is to the south.

Introduction

North of Bathurst the post-Taconian sediments of mainly Silurian age start with the basal late Llandovery Weir (Armstrong Brook) Formation resting unconformably on Ordovician rocks of varying ages. They are immature volcano-sediments of mainly alluvial fan origin deposited in separate fault-bounded basins and show rapid thickening, facies variations, and divergent provenances. They are also associated with some volcanic activity. By late Llandovery to Ludlow time the sediments consist largely of near shore to outer shelf mixed terrigenous and carboniferous facies representing a number of large-scale transgressive-regressive sequences. By Wenlock time volcanic activity intensified and continued sporadically throughout the rest of the Silurian and into the early Devonian.

The Silurian succession at Quinn Point (Fig. 2) is one of the most complete in the region and has been subdivided into a number of zones or biofacies on the basis of lithology and fauna. On the beach south of the road is a continuous section, about 700 m long, of red sandstones and grits of the Weir Formation which grades into the nodular siltstones and then limestones of the La Vieille Formation at Quinn Point proper. To the north are folded conglomerates with basalt clasts from contemporaneous basalt flows, nodular and fossiliferous limestones, and sandstones and siltstones of the Weir Formation with minor amounts of La Vieille limestone in the core of a syncline. About 350 m to the north the sequence is topped by the Carboniferous unconformity.

Quinn Point

The Weir Formation begins with a dark-red and maroon coarse sand facies that is texturally and mineralogically immature and very ferruginous. It becomes finer upsection, while lower downsection is associated with volcanics (see Fig. 2). They show large scale cross-bedding, debris flows and other characteristics of alluvial fan continental-type sedimentation. There are no fossils present in these beds. In contrast to this, in the north in equivalent sediments, there are Eocoelia beds indicating a delta front coastal environment.

Above these red beds are intermittently exposed interbedded sandstones and limestones of the Weir and La Vieille Formations. From the oldest to the youngest, the following biofacies have been established (Fig. 3): Dolerorthis, Tabulate coral, Rhynchonellid, Eocoelia, and Paleocyclus-Atrypa Biofacies of the Weir and Costistricklandia, Stromatoporoid, Leptaena-Protochonetes and Algal Biofacies of the La Vieille Formation. On the basis of sediment type and faunas the first four biofacies are interpreted as shallow water nearshore communities followed by a deep water environment community (Paleocyclus B.). Above this biofacies, heavy horizontal bioturbation, common erosional features and nodular limestone all point to an outer shelf environment which is followed by gradual shallowing during the Stromatoporoid to Algal B. deposition. Overall the sequence is one of transgression followed by regression with the deepest water depth being achieved during the Costistricklandia B. deposition.

Source

Fyffe, L.R. and Noble J.P.A., 1985. Stratigraphy and Structure of the Ordovician, Silurian and Devonian of Northern New Brunswick. Geological Association of Canada, Mineralogical Association of Canada, Joint Annual Meeting, Fredericton '85, 56p.

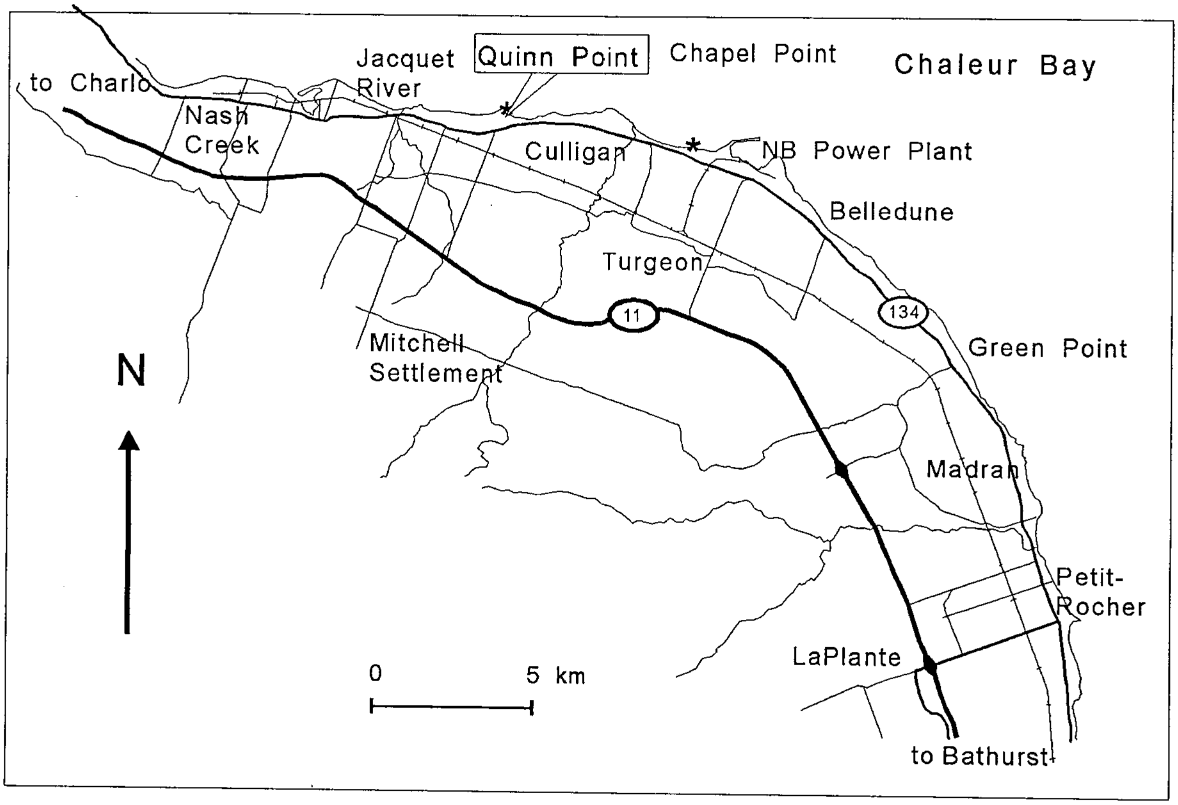

Figure 1: Location map showing Quinn Point

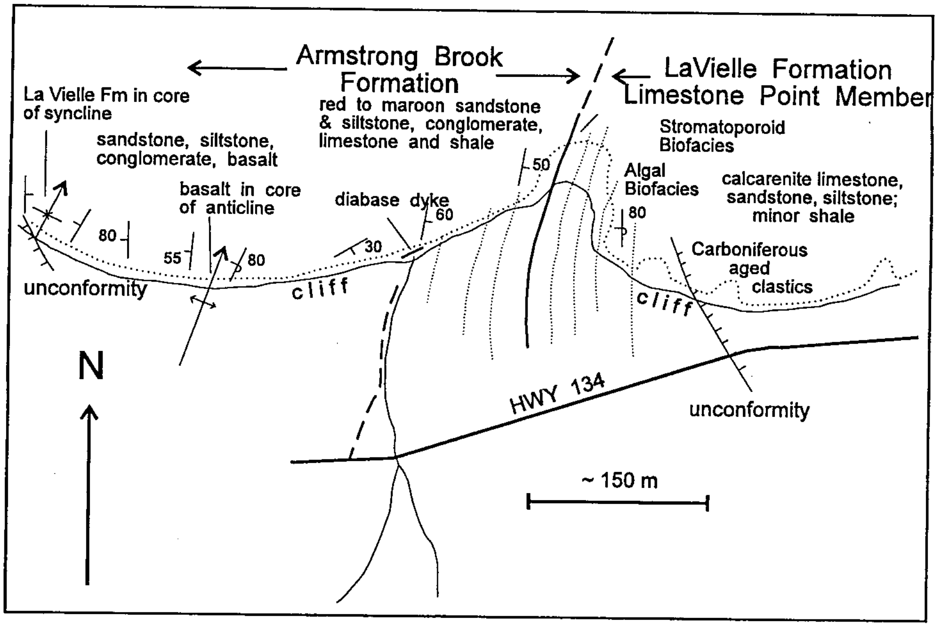

Figure 2: Geological map of the Quinn Point area. Modified after Fyffe et al., 1985 and Howells, 1975.

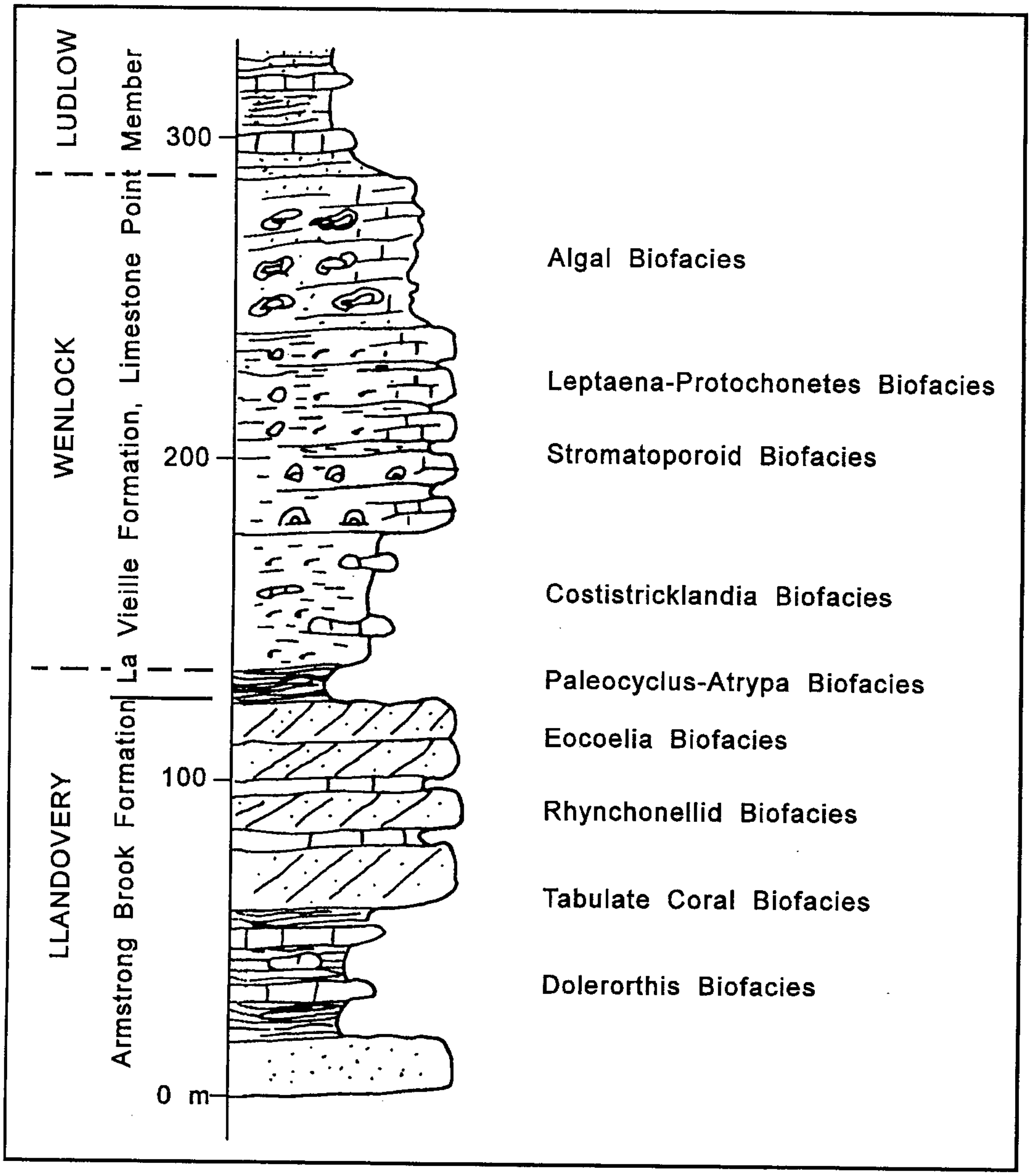

Figure 3: Generalized stratigraphic section at Quinn Point. Modified from Fyffe et al., 1985.

Highway 11

Chaleurs Group Basalt and Rhyolite along Highway 11

Purpose

Along Highway 11 are superb examples of felsic and basaltic volcanism of the Chaleurs Group.

Location

The outcrops are north of Bathurst on Highway 11, 10.3 km north of the bridge over highway 357 or 7.5 km south of Charlo (Fig. 1). About 1 km north of the Benjamin River bridge are a series of outcrops on either side of the road stretching north for approximately 2 km. The basalt exposures are to the south on the east side of the road and the best flow banded rhyolites are about 500 m along on the west side. Rhyolites dominate most of the exposure.

Introduction

During the late Silurian to early Devonian (Wenlockian and younger) in northern New Brunswick volcanic activity intensified from the sporadic episodes that are typical of the Llandoverian. Composition varies from basalt to rhyolite and alteration has greatly affected their geochemistry so affinities are not known. The volcanics are associated with minor sediments and volcani-clastics. The Benjamin Formation is the youngest of the Chaleurs Group and is folded about northeast-southwest plunging synclines.

Description

The basalt flows are fairly typical vesiculated to massive basalts with plagioclase phenocrysts and weathered zones between flows (Fig. 2). They are separated from the rhyolites to the north by a postulated fault. The rhyolites are purplish to green, massive, flow layered, porphyritic and/or pyroclastic. Phenocrysts are generally small, 1–2 mm, subhedral to round K-feldspars and are never more than a few percent by volume. The pyroclastic units have small fiamme or flattened pumice lapilli, as well as the occasional rounded lithic flow layered and/or porphyritic clast or lapilli. Flow layers in both the pyroclastic and massive units are folded and contain minor thrust ramps reminiscent of lateral movement after deposition. The flows are cut by small vertical felsic dikes up to 5 m thick exhibiting chilled margins.

Underneath the volcanic pile are volcani-clastic sandstones and conglomerates showing grading, imbrication, cross-bedding, and erosional channels indicating the pile is right way up. Beds are variably thick from 3 to 30 cm with the occasional larger one. Rounded red rhyolite clasts are prominent throughout the sediments.

Bedding and flow layering generally are shallow to moderately dipping. Overall the whole succession is folded into west-plunging open folds with wavelengths of approximately 500 m.

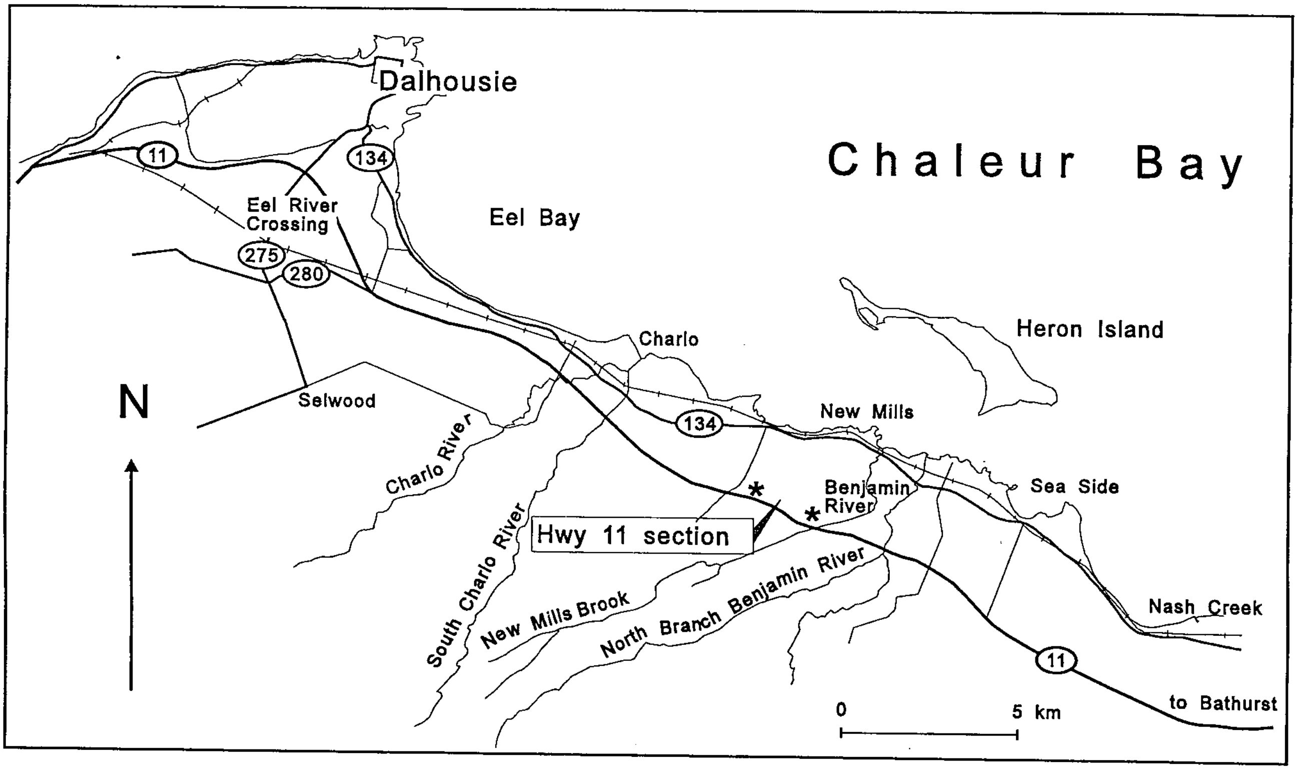

Figure 1: Location map for Highway 11 volcanic rocks at New Mills Brook.

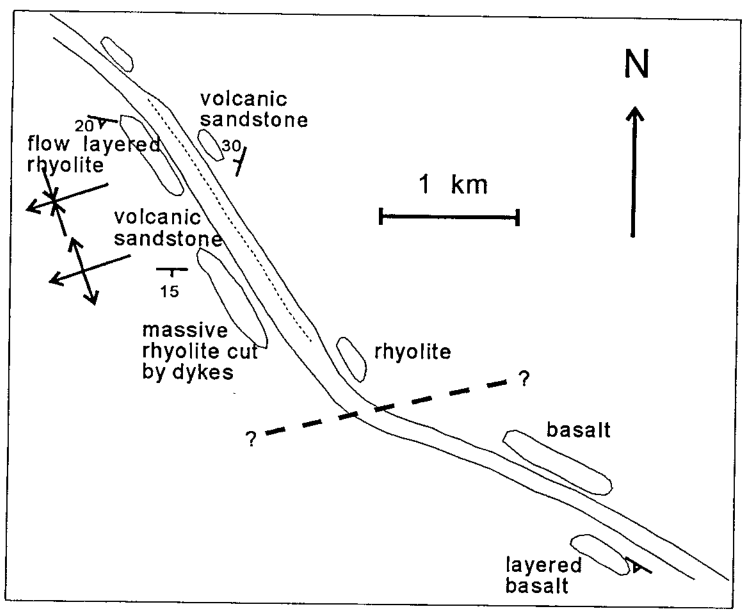

Figure 2: Detailed geology of volcanic rocks exposed along Highway 11 north of bridge over New Mills Brook.

Dalhousie—Eel Bay

Sedimentary and Volcanic Rocks of the Lower Devonain Dalhousie Group

Purpose

The Eel Bay section contains good exposures of the Lower Devonian Dalhousie Group containing calcareous siltstone, shale and limestone interbedded with mafic volcanic flows and palagonite (hydrated and devitrified basaltic glass) tuffs.

Location

Within Dalhousie the top of the section can be reached along the shore at low tide by going south from the town park (Fig. 1). Alternatively the bottom of the section can be accessed through the NB Power plant at Darlington on Eel Bay.

Introduction

The rocks south of the Restigouche River and within the Lower Devonian Dalhousie Group are an intimate succession of interbedded sediments and volcanic rocks. They are best exposed along the coast where relationships can be seen; within the bush only the harder volcanic rocks crop out. The rocks rest conformably on the Silurian and older strata and are divided into two units. The lower unit is chiefly sediments with minor amounts of volcanics and the upper unit is a thick succession consisting largely of volcanic rocks with local sediments found in thin bands.

The Dalhousie section is very accessible and well exposed. The sedimentary rocks contain an abundance of fossils that have been correlated with rocks elsewhere (other parts of eastern North America as well as Europe), of similar age, and indicating they are contemporaneous and within the same depositional basin. The lowermost and upper volcanic rocks likely originated from a volcano that occupied the site of the present day Dalhousie Mountain. The volcanics in the middle of the section came from a different source to the east or northeast.

Description

The succession begins on the shore near Darlington with calcareous shale beds (Fig. 2), labelled

Zone	*Description*	*Thickness*	
		feet	*metres*
	andesite (4th and 5th flows	40+	12+
	intrusive andesite and breccia	90	27
	andesite (1st to 3rd flows)	85	26
	gap in section	250	76
16	arenaceous limestone with grey shale	25	7
15	grey limestone	2	0.6
14	thin-bedded shale with limestone	35	11
13	tuff	1	0.3
12	blocky calcareous shale	2	0.6
11	tuff with thin limestone and shale	30	9
10	thin grey shale	10	3
9	limestone and grey shale	75	23
8	calcareous sandy shale	20	6
7	arenaceous limestone	7	2
6	arenaceous limestone with agglomerate and andesite	280	85
5	grey limestone	10	3
4	tuff	12	4
3	grey arenaceous shale with limestone	40	12
2	grey calcareous shale with limestone	125	38
1	silicified limestone with shale	30	9
	basaltic tuff and breccia	30	9
	basalt	15	5
	palagonite tuff	180	55
0	calcareous shale	90	27

Zone 0 by Alcock (1935), and dips moderately northeastwards towards Bon Ami Point. The total section is in excess of 1384 feet (422 m) with increasing volcanism upsection. The following description and table is summarized from that source:

The lowermost beds (Zone 0) are thin-bedded calcareous shale, fine grained sandy shale and shaly limestone containing a few brachiopod fossils. Overlying this is the dark green to greyish mafic hylaclastite (palagonite tuff) that is well bedded but highly weathered, soft and crumbly to the touch. Above this is a single amygdaloidal basalt flow followed by volcanic tuff and breccia closely resembling the underlying flow. All these volcanics are believed to have emanated from the volcanic centre that lies behind the town of Dalhousie and from an early more basic phase of the more volumetric andesitic flows that follow later.

The intervening sedimentary rocks (Zones 1 to 6) were deposited directly on top of the volcanics and contain abundant corals near their base and brachiopods and pelecypods further up. It is unfossiliferous at the top. Zone 6, a fossiliferous limestone, is exposed intermittently under the major andesitic flows and agglomerates and strikes into them to the north of a small fault at the point. Alcock (1935) suggests that there was a small interval of time between the deposition of Zone 6 and the andesite flows which resulted in their partial removal.

The volcanics overlying the lower member form a steep cliff, 30–40 metres high that extends 300–350 metres along the shore. The lower portion is a dark green agglomerate composed of irregular shaped and variously sized bombs up to 60 cm diameter as well as smaller clasts. Above this is a dense, amygdaloidal andesitic lava flow cut by numerous calcite veins. Tracing these flows to the west shows them not to be associated with the vent at Dalhousie Mountain but probably a vent further to the east. This along with the erosional surface at their base prompted Alcock (1935) to separate the two sequences of Dalhousie beds into lower and upper members.

Above these volcanics are the second sequence of sediments (Zones 7 to 16). The lowermost limestone rests conformably on top of the andesite even though the surface is very irregular. All units are fossiliferous with fossils ranging from corals, pelecypods and brachiopods in the lower zones to mainly brachiopods and pelecypods with few other fossils in the middle zones to abundant corals and few shelly fauna in Zone 16. Exposures above Zone 14 are generally poor with the finer grained beds not exposed.

Above Zone 16 there is a space of 250 feet (76 m) of cover before andesite crops out again. Above this interval five pyroxene andesitic flows with associated volcanic breccia and agglomerate can be recognized with the lower flows cut by andesitic dikes feeding the upper flows. The flows are variably massive to amygdaloidal with vesicles filled by calcite, chlorite or limonite; fractures are filled with chalcedony and agate. Most fractures are short and narrow but many are over 20 m long and several centimetres thick with banding and stains of red or violet. Many of the fractures in the lower three flows are truncated by the upper flows indicating there was a period of fracturing prior to their eruption. Further faulting followed these flows. The section continues to the north but much of it is covered and only the more resistant andesites are exposed.

Much of the town of Dalhousie, as well as Dalhousie Mountain, is underlain by pyroxene andesite, many outcrops can be seen around the town and along the shore. This mountain is the truncated core of a volcano and represents the last phase of volcanic activity in the area that fed these latest flows. The islands several hundred metres offshore have a similar composition.

Source

Alcock, F.J. 1935. Geology of Chaleur Bay Region. Canada, Department of Mines, Bureau of Economic Geolgy, Geological Survey Memoir 183, 146p.

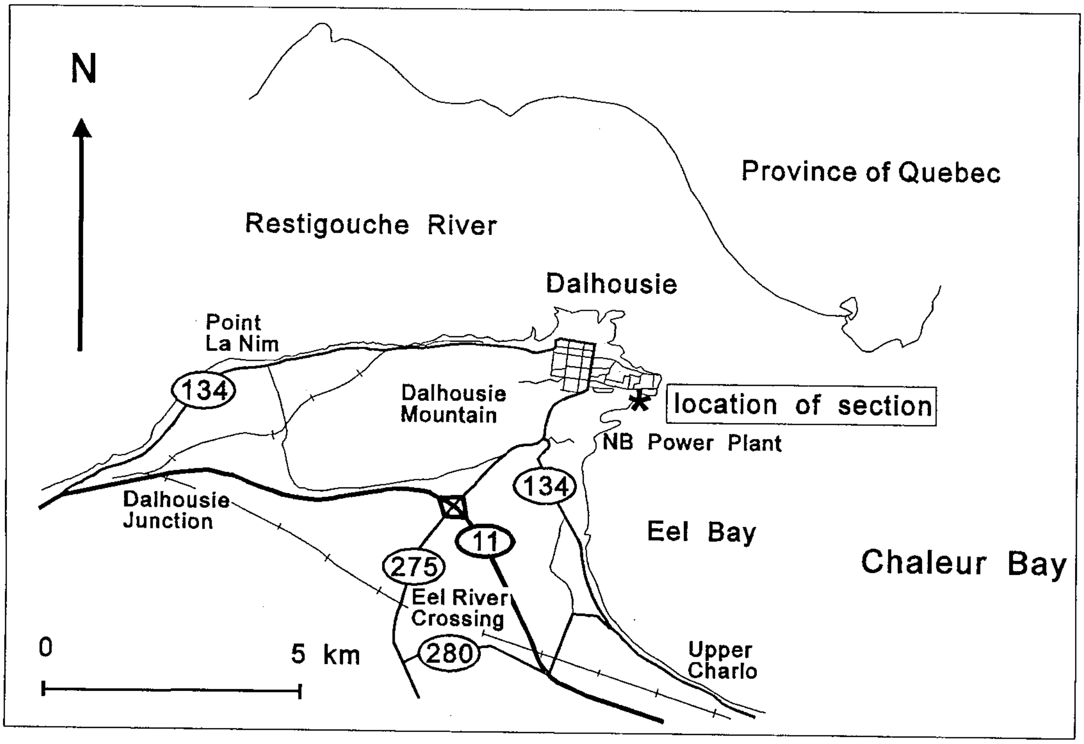

Figure 1: Location map of the Eel Bay section

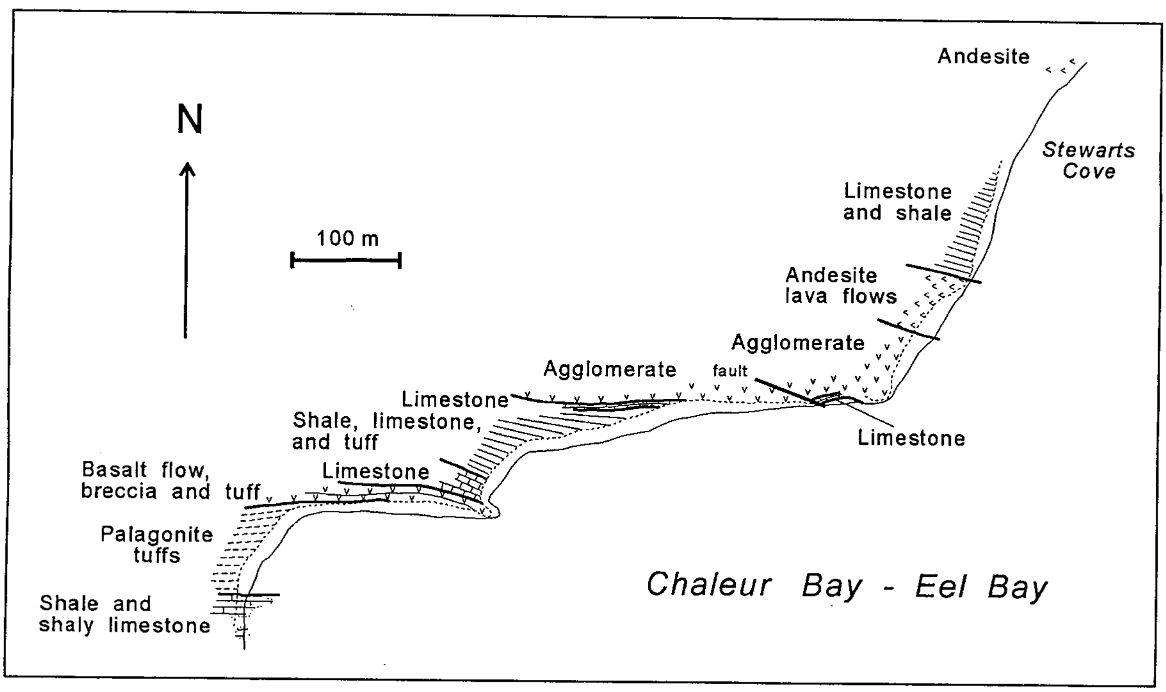

Figure 2: Detailed geology along north shore of Eel Bay at Dalhousie. Modified from Alcock, 1935.

Other Sources of Information — Books, Videos, Maps

General Books and Publications

Field Guides of Rocks and Minerals of North America

Dictionary of Geology (Bantam, Penguin, or American Geological Institute (AGI))

Be Expert with Map and Compass by B Kjellstrom

Safety Manual: Mineral Exploration in Western Canada, BC Chamber of Mines

First Aid Manuals

Geologically related Videos

Atlantic Geoscience Society: *The Mineral Wealth of Atlantic Canada, The Appalachian Story, The Recent Ice Age, Offshore Oil and Gas*

also others available from :

Department of Education, Education Media Services, Provincial

Energy, Mines and Resources, Communications Branch, Federal

National Film Board of Canada

Public Archives, Provincial

Teaching Modules and other programmes

Scientists in the Schools (SITS) —Nova Scotia Centre for Marine Geology, Dalhousie University Halifax. (902) 494-6785

Panning for Gold — held in the spring of each year contact Department of Natural Resources, Nova Scotia

Prospecting and Geology — held in the fall or spring of each year. Contact local Departments of Natural Resources

Evening courses in geology and related subjects at the introductory level at some universities and colleges

Festivals and similar events

Canada's Parks Day — held in mid-July, outdoor recreational festival, guided tours, etc.

Parks are for People — walking tours of selected parks and areas at Peggy's Cove, Minas Basin, Five Islands, Arisaig, and Taylor's Head co-ordinated through Nova Scotia's provincial parks

Rock Hound Round Up — Parrsboro's annual event in August of each year. Contact Parrsboro Museum. (902) 254-3814

Educational Kits

Rock, Mineral and Fossils kits are available for loans to teachers from the Nova Scotia Museum, Halifax and The New Brunswick Museum, Saint John.

Rock and Mineral Sets — Geological Survey of Canada sets of 21 rocks and 21 minerals or 36 rocks and minerals for sale. Energy, Mines and Resources, Communications Branch Ottawa. (613) 992-8163

Museums with Geoscience sections

Cape Breton Centre for Heritage and Science, Sydney, NS

Fundy Geological Museum, Parrsboro, NS

Inverness Miner's Museum, Inverness, NS

Londonderry Mines Museum, Londonderry, NS

Marble Mountain Library & Museum, West Bay, NS

The Miner's Museum, Glace Bay, CBI

Moose River and Area Museum, Moose River, NS

Nova Scotia Museum, Main Branch, Halifax, NS

Nova Scotia Museum of Science and Technology, New Glasgow, NS

Springhill Miner's Museum, Springhill, NS

Wallace Bicentennial Museum, Wallace, NS

Joggins Fossil Centre, Joggins, NS

Ovens Natural Park, Riverport, NS

The New Brunswick Museum, Saint John, NB

The Grindstone Museum, Sackville, NB

Grand Falls Information Bureau, Grand Falls, NB

Restigouche Regional Museum, Dalhousie, NB

New Brunswick Mineral and Mining Interpretation Centre, Petit-Rocher, NB

Grand Manan Museum, Seal Cove, Grand Manan Island, NB

The Newfoundland Museum, St. John's, NF

Baie Verte Peninsula Miner's Museum, NF

Mary March Regional Museum, Grand Falls, NF

St. Lawrence Miner's Museum, St. Lawrence, NF

Newfoundland Freshwater Resource Centre, St. John's, NF

Labrador Heritage Museum, Happy Valley-Goose Bay, Labrador

Geological Displays

Acadia University, Nova Scotia

Atlantic Geoscience Centre — Bedford Institute of Technology, Dartmouth

Dalhousie University, Halifax

Mount Allison University, Sackville, NB

Saint Mary's University, Halifax

St. Francis Xavier University, Antigonish

DalTech (formerly Technical University of Nova Scotia), Halifax

University of New Brunswick, Fredericton

Memorial University of Newfoundland, St. John's

Libraries

Universities and colleges including:-

Nova Scotia Community Colleges

University College of Cape Breton, Sydney

University of King's College, Halifax

Nova Scotia College of Art and Design , Halifax

Mount Saint Vincent University , Halifax

University of New Brunswick, Saint John

University of Moncton, Moncton

University of Prince Edward Island, Charlottetown

West Viking College, Stephenville

Magazines and Newsletters that might be of interest

BIOME — a publication of the Canadian Museum of Nature

Discover, The World of Science

Earth, The Science of Our Planet

Earthquakes and Volcanoes (A US Geological Survey publication)

Equinox

Geos (An Energy, Mines and Resources, Canada publication)

Geotimes (American Geological Institute)

Journal of Geoscience Education

Museum News (Nova Scotia)

National Geographic

Oyez (The Royal Society of Canada publication)

Prospectives (joint Federal and Nova Scotia government publication)

Wat on Earth (University of Waterloo publication)

Maps — from Government Bookstores, selected bookstores and sporting goods stores

Geological maps of Nova Scotia, New Brunswick, PEI, and Newfoundland

Geological Highway Map of Nova Scotia

Natural History Map of Nova Scotia

Map book of the Province of Nova Scotia

Geological Highway Map of New Brunswick and PEI

Newfoundland and Labrador: Traveller's Guide to the Geology and Guidebook to Stops of Interest

Publications — specific reports, maps, field guides, walking tours, etc.

Atlantic Geoscience Society, c/o Dalhousie University, Dept. of Earth Sciences

Chamber of Mineral Resources of Nova Scotia

Department of Natural Resources, Minerals and Energy Branch, Nova Scotia

Newfoundland Department of Mines and Energy, Geological Survey Branch

Newfoundland Department of Mines and Energy, Petroleum and Energy Resources Branch

New Brunswick Department of Natural Resources, Geological Survey Branch

New Brunswick Department of Natural Resources, Energy Branch

Prince Edward Island Department of Energy and Forestry

Department of Natural Resources, Canada

Environment Canada

Fisheries and Oceans, Canada

The Museum of Nova Scotia

The New Brunswick Museum

The Newfoundland Museum