

Department of Earth & Environmental Sciences

NOVA SCOTIA FIELD GUIDE



Arthur D. Storke Memorial Expedition



August 23 to September 2, 2012

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Front Page: Nova Scotia Field Trip Participants at Cape Breton Highlands National Park

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I. Preface

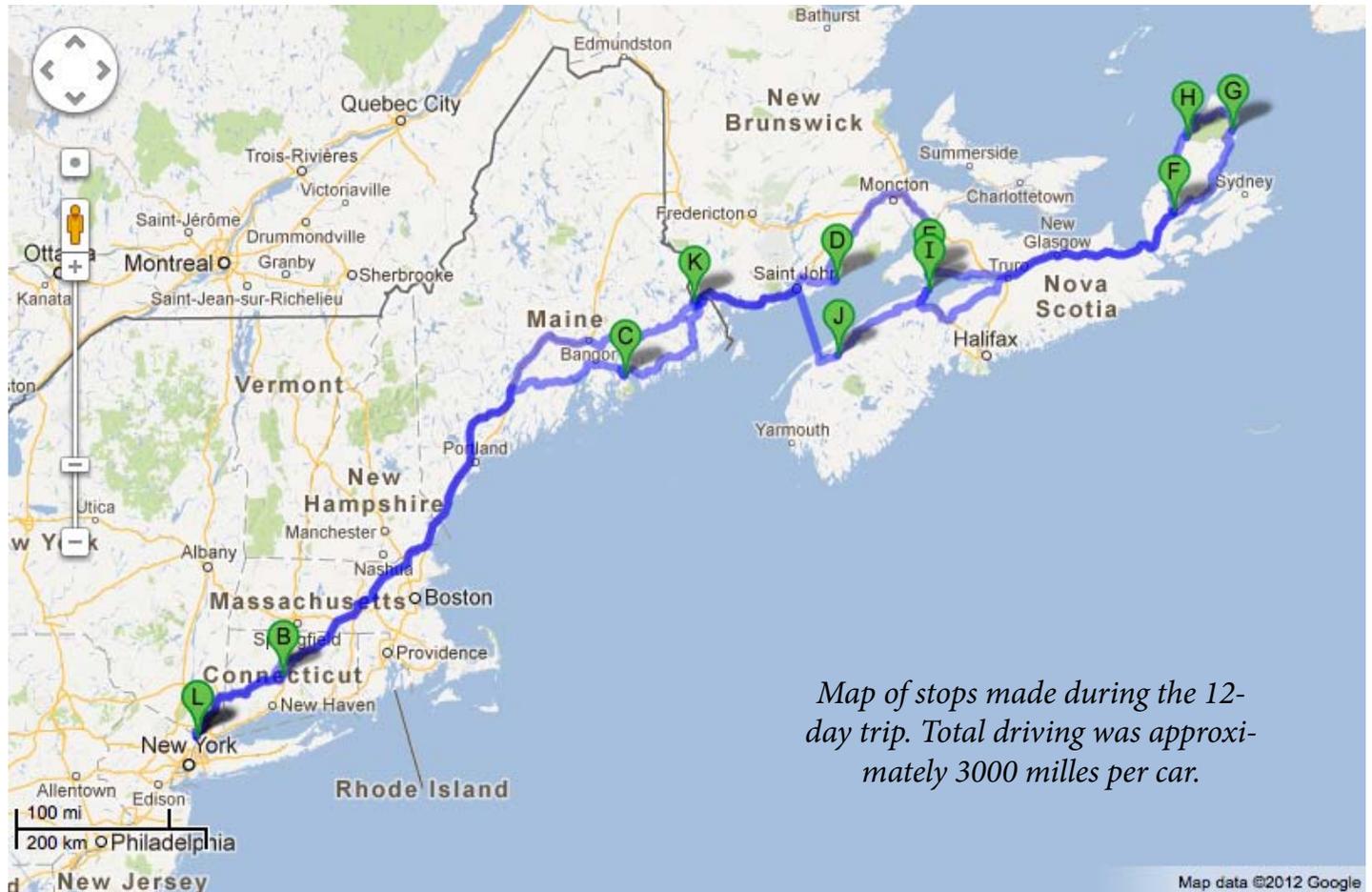
It's not that far away.

From the vantage point of New York City, Nova Scotia seems like another world. But a couple of rental cars and a full day of driving from New York will bring you to the doorstep of Nova Scotia, where amazing outcrops representing the full secession of Phanerozoic tectonic, climate, and evolutionary history are preserved among the grandiose vistas and welcoming hospitality of Atlantic Canada's most populous province. Plus, the Mesozoic history of the opening of the Atlantic Ocean, recorded in the sediments of the Hartford Basin, and the Quaternary glacial geology of eastern Maine just happen to be convenient stops along the way.

That, in a nutshell, provided the initial motivation for the trip on which this guide is based. Over the course of 2011 and 2012, a group of Ph.D. students at Columbia University's Department of Earth and Environmental Sciences representing a diversity of sub-disciplines--including geodynamics, stratigraphy, sedimentary geology, igneous petrology, glaciology, geochemistry, oceanography, and paleoclimate--organized an eleven-day trek across Nova Scotia for late August, 2012. A weekly, student-led seminar series met during the month of June, where presentations from field trip participants on specific geologic topics or locations of interest drove our intellectual preparation for the field excursion. Many of the specific topics or locations discussed in the seminar were visited during the course of the trip and are covered in the Geologic Overviews.

This guide was prepared following the 2012 Arthur D. Storke Memorial expedition to Nova Scotia, with broad overviews of Nova Scotian geology and detailed itineraries and descriptions of individual outcrops visited. Our hope is that this guide can serve as a blueprint for future trips to Nova Scotia. For faculty and students in the northeastern United States, outstanding outcrops detailing the Phanerozoic history of continental collision and rifting are not that far away. Nova Scotia is waiting.

II. Trip Itinerary



DAY 1 (August 23): Leave New York (stop L), travel to Lemoine, ME (~9 hour drive). Stay at Lemoine State Park (stop C). Make stops at outcrops of the Hartford Basin in CT (stop B):

- Road cut showing thick sequence of pillow basalts from the Talcott Basalt
- Outcrop showing Milankovitch cycles in lacustrine beds of the East Berlin Fm.
- Dinosaur State Park (dinosaur tracks)
- Portland Fm. arkosic sandstone

DAY 2 (August 24): Leave Lemoine and make stops along the Ice Age Trail (<http://iceagetrail.umaine.edu/>). Arrive in St. John, New Brunswick, Canada (6hr10), and camp at St Martins Beach (stop D), at the Seaside Campground, ~ 1hr north of St. John.

DAY 3 (August 25): Leave St John area and arrive in Parrsboro, Nova Scotia (5hr20). Camp at Five Islands Provincial Park (stop E). Make stops at Fundy Basin outcrops:

- Honeycomb Pt. Fm at St. Martins Caves
- Wolfville Fm. at Melvin Beach
- Albert Fm. along a roadcut at km 184 on Rt. 1.

DAY 4 (August 26): Depart from and return to Five Islands Campsite (stop E).

- Visit to Joggins Fossil Cliffs (<http://jogginsfossilcliffs.net/>)
- Cape Chignecto Park: Blomidon Fm and Cobequid Fault Zone
- Clarke Head: Blomidon Fm, North Mountain Basalt and Windsor Group

Trip Itinerary (*continued*)

DAY 5 (August 27): Depart from and return to Five Islands Campsite.

- Visit to Fundy Geological Museum (<http://museum.gov.ns.ca/fgm/en/home/default.aspx>)
- Wasson Bluffs: Minas Fault Zone
- Five Islands Provincial Park: Blomidon Fm, McCoy Brook Fm, and North Mountain Basalt (<http://parks.gov.ns.ca/parks/fiveislands.asp>)

DAY 6 (August 28): Leave Parrsboro and arrive in Antigonish, NS (2hr35). Meet with Mike Melchin, St. Francis Xavier University. View Ordovician-Silurian outcrops, brachiopod paleontology, and structures related to the Avalon-Meguma suture. Depart Antigonish ~5pm and drive to Whycomomagh campsite (1hr20min, stop F). Eat dinner at Brass Door Restaurant in Whycomomagh.

DAY 7 (August 29): Loop around Cape Breton Highlands National Park (stops **G thru H**, Cabot Trail, <http://travel.nationalgeographic.com/travel/road-trips/cape-breton-nova-scotia-road-trip/>), before returning to Whycomomagh campsite. Total loop, with stops, takes about 7 hours. Eat dinner at Coal Mine Café in Inverness, NS.

DAY 8 (August 30): Leave Whycomomagh, arrive at Cape Blomidon Provincial Park in Wolfville, NS (stop **I**, 4hr30 drive) making stops along the way:

- Coal Mine Point, Finlay Harbor, Mabou NS (salt diapir)
- Shubenacadie Canal, rt. 236
- Burntcoat Head Park

DAY 9 (August 31): Stay in Wolfville area at Cape Blomidon Provincial Park. Check out Bay of Fundy Tides at Delhaven Harbor, scout for dinosaur tracks and fossils near Paddys Island in North Medford, and view North Mountain Basalt outcrops near Cape Split.

DAY 10 (September 1): Spend the morning in Blomidon Provincial Park trails, then leave Wolfville towards Digby, NS (stop **J**, 1hr45). Catch 4pm ferry from Digby, NS to St John, NB. The ferry ride is 3-4 hrs long (it is only 45 miles). After we arrive in St John, NB, drive across the border and stay in Calais, ME at the Boidi Motel (stop **K**, 1h30 drive).

DAY 11 (September 2): Leave Calais and drive directly back to Palisades, NY (9hr).

III. Introduction to Nova Scotia

contributed by Alexander Lloyd

When you think of Nova Scotia, what comes to mind?

Whether it's the seafood industry, its venerated history of shipbuilding, the cities of Halifax or Sydney, or its standing as one of the three Canadian Maritime Provinces (unofficially Canada's Ocean Playground), people generally associate Nova Scotia with the Atlantic Ocean in one way or another. However, the field trip that is outlined in the following guide is decidedly not maritime in nature and highlights what else Nova Scotia has to offer.

Nova Scotia is a Canadian province fronting the Atlantic Ocean. Though small in size in comparison to its fellow provinces, it boasts a complex climate and geography. Its provincial capital, Halifax, is located approximately half way between the equator and the North Pole. It has a unique climate that is best described as mid-temperate and continental yet buffered from extreme temperatures by the ocean. The field trip participants found the weather very pleasant for late summer and some were even disappointed that there wasn't a reprieve from the dry heat waves experienced by the greater New York area all summer.

Initially discovered by an Italian who was hired by the British crown, Nova Scotia (literally translated as New Scotland) has had a long history of colonization and national affiliation. As a result, there are unique, culturally isolated groups (Gaelic, Acadian, and Indigenous) scattered throughout the province. Originally populated by the Mi'kmaq, French colonists were the first to establish a colonial government in the territory. After the Treaty of Utrecht in 1713, English colonialists took control of Acadia (Nova Scotia minus Cape Breton). By 1755, the majority of the French population was expelled and replaced with New England Planters. During the American Revolution, there was a possibility that Nova Scotia would join the war against Britain as "the 14th American Colony," but the disproportionate number of loyalists and economic disruption caused by American privateers effectively ended the prospect of statehood. During the first Canadian federal election in 1867, Nova Scotia elected a significant number of anti-Confederate representatives that favored self-governance for Atlantic Canada. Without a majority in the national government and with no support from Great Britain, Nova Scotia was not successful in its ambitions to secede from the newly formed Canadian confederation.

In many ways, Nova Scotia really is "New Scotland." The largest ethnic group in the province is Scottish, which accounts for 28.3% of the population. The only Scottish-style single malt whiskey distillery in Canada can be found on Cape Breton Island. When driving through the town of Antigonish, one observes street signs in both English and Gaelic. Though on opposite sides of the Atlantic, the Scottish Highlands and Cape Breton Island share not only a similar appearance, but also a shared geologic history. Both mountain chains were formed during the Caledonian/Acadian orogeny in the Early Devonian.

Though the economy of Nova Scotia has historically been labeled as natural resource based, new challenges in the 21st century (the collapse of the cod stocks on the Scotian Shelf and cessation of coal mining on Cape Breton) have required the province to diversify. Tourism and competitive shifts in the manufacturing sector have improved the state of the historically resource-based economy. Though not considered a significant exporter by volume, the province is the world's largest exporter in certain niche industries including Christmas trees, lobsters, and wild berries.

Nova Scotia is not only unique and diverse in its historic and cultural characteristics, but as you will discover in the following field guide, there are a multitude of earth science subjects that can be explored in Canada's second smallest province. Whether you are interested in coastal processes, Paleozoic tectonics, or dinosaurs, Nova Scotia offers something for every earth scientist.

(All information from Wikipedia)

IV. Geologic Overviews

Paleozoic History of Nova Scotia

Contributed by Marc Vankeuren

The Geology of Nova Scotia is a diverse mosaic of geologic history representing much of the Phanerozoic and is greatly exposed along the shorelines. On a broad scale, Nova Scotia is divided into two major geological regions, the Meguma Terrane and the Avalon Terrane (Fig. 1) (Louden, 2002). Terranes represent microcontinents, which have been amalgamated over geologic time into a single landmass. The Avalon Terrane occupies the northern and eastern portions of the Nova Scotia. The Avalon Terrane was formed as a volcanic arc at the mid-ocean ridge of the proto-Atlantic Ocean, named the Iapetus Ocean, during the Cambrian Period. At this time the Avalon Zone was part of the northern European sub-continent called Avalonia. In the Early Ordovician, this landmass was far south of the equator. During the Late Ordovician Period (Fig. 2), Avalonia collided with, and amalgamated to Laurentia, closing the Iapetus Ocean - a body of water that separated Laurentia from Avalonia. These units consist of dark green basaltic lava flows overlain by orange-red felsic lava flows (rhyolites), felsic ash flows and tuffs. The basalts show columnar jointing and gas cavities either filled or open at the tops and bottoms of the flows. Red to purple gravelly layers occur between the basaltic lava flows and are interpreted as ancient soils, which developed between volcanic eruptions (Fossils Of Nova Scotia, 2012).



Figure 1: Map of Nova Scotia showing two major Geologic Terranes. From Nova Scotia Museum, *Fossils of Nova Scotia*. SchoolNet Digital Collections. <http://museum.gov.ns.ca/fossils/index.htm>.

The Meguma Zone differs from the Avalon Zone because the two areas were widely separated and in different environments. In the Cambrian Period, the Meguma Zone lay seaward of a large continental landmass, believed to have been North Africa. The Rheic Ocean separated Paleo North America (Laurentia) from Paleo -Africa, -South America, -Australia, -Asia and -Antarctica (Gondwana). The Rheic Ocean separated the two landmasses of Laurentia (now including Avalonia) and Gondwana. Large volumes of sediment were transported from the African continental shelf of the landmass

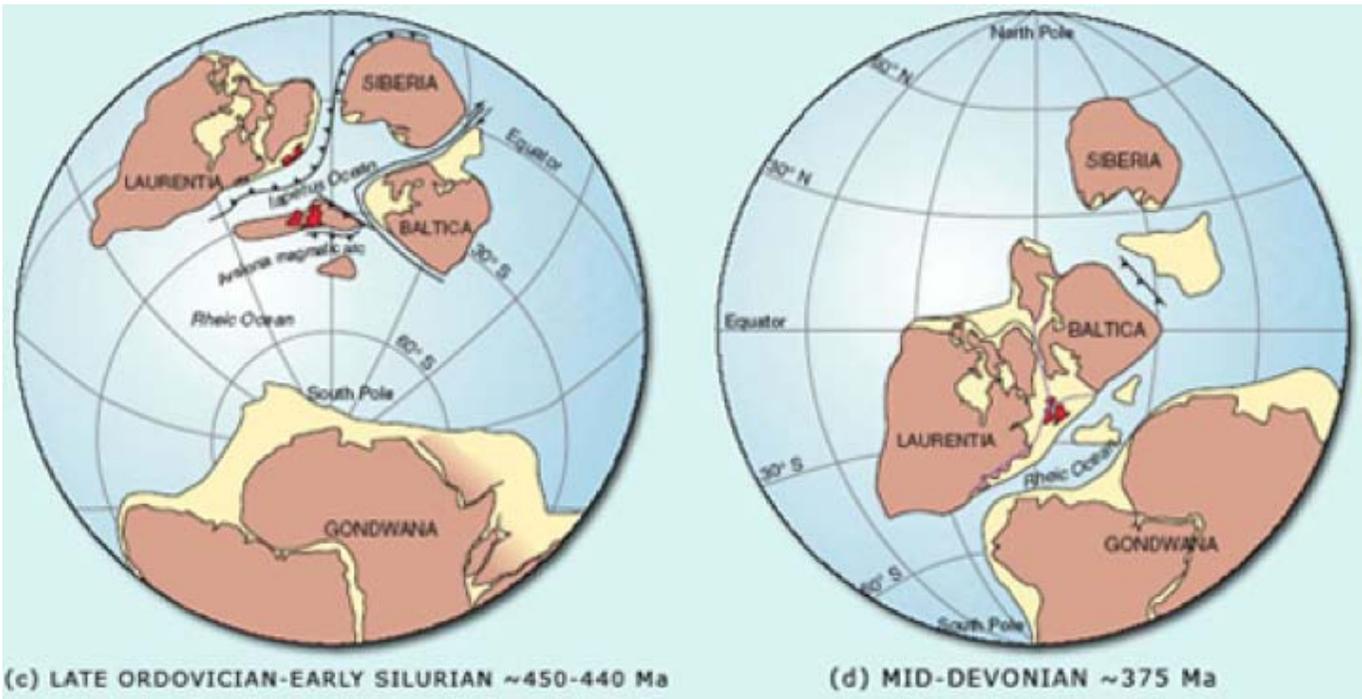


Figure 2: Paleogeographic map from Ordovician to Middle Devonian. From "A global view of Earth history" - Geological processes in the British Isles. The Open University. <http://openlearn.open.ac.uk/mod/oucontent/view.php?id=398700§ion=3>

to the depositional area of the Meguma Group by turbidity flows. These deposits are represented by the thick, uniform sandstones of the Goldenville Formation and are overlain by slates of the Halifax Formation (White, 2008).

Before the Devonian, the Avalon and Meguma Zones of Nova Scotia were widely separated because they belonged to different continental landmasses (Laurasia and Gondwanaland, respectively). The distance between the two continents was substantial in that Laurentia was situated just south of the equator under warm tropical climates while Gondwana experienced mostly cold climates, including extensive glaciers. The formation of Pangea, however, joined them together for the first time to form the single geological entity we now know as Nova Scotia. As the Rheic Ocean continued to close and Gondwanaland migrated northward, eventually suturing into the Avalon Terrane forming the New Brunswick Fracture that runs through the middle of Nova Scotia. From the Middle Devonian on, the province was in one piece and had a similar geological history continuing to the present day (Fossils Of Nova Scotia, 2012).

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Mesozoic History of Nova Scotia

Contributed by Raj Moulik

The Mesozoic Era, spanning between 252-66 Mya, is marked by the formation of rift basins and the predominance of reptiles in Nova Scotia. Prior to the Mesozoic, Nova Scotia occupied a central position in the supercontinent Pangea and was tectonically quiescent. A new phase of crustal motion started in the Triassic that eventually resulted in the breakup of Pangea and the formation of the Atlantic Ocean. The breakup was marked by the formation of an enormous series of rift basins along a wide zone covering the present-day Gulf of Mexico, Nova Scotia, Morocco and much of western Europe. The Central Atlantic Margin (CAM) component of these rifts includes the exposed remnants of the Newark Supergroup in eastern North America (Figure 1). The rifting was accompanied with basaltic flows and sedimentary infill into basins that were the precursors of the various modern coastal features such as the Bay of Fundy and the Minas Basin. Since the continental half-graben and half-graben complexes got filled with kilometers of such sedimentary and igneous strata, the resulting sequences provide a comprehensive record of early Mesozoic climate, tectonics and evolutionary patterns. This synopsis outlines the major features of these sequences, their relation to overall tectonics and their characteristics in the Fundy Basin.

The prominent feature of the CAM basins is the cyclicity in their lacustrine sequence, which allows quantitative analysis of the intertwined effects of tectonics, plate position, and climate over a large latitudinal spread of about 30 degrees in paleolatitude and from the Late Triassic and to the earliest Jurassic. This sedimentary cyclicity was primarily controlled by lake-level variations following quasi-periodic Milankovitch-type climate cycles (Hays et al., 1976) and their climatic significance has been extensively studied in the CAM basins (e.g. Smoot, 1991). The fundamental sedimentary cycle in these basins is a transgressive-regressive lacustrine sequence termed a Van Houten cycle (e.g. Olsen, 1986). The vertical sequences outline three orders of modulating cycles that are related to subaerial exposure or deposition in deep water within successive Van Houten cycles. The frequency analysis of these sections reveal that these the climate changes that controlled the lake levels were controlled by the precession cycles and further modulated by the eccentricity cycles of the Earth's orbit (Olsen, 1986).

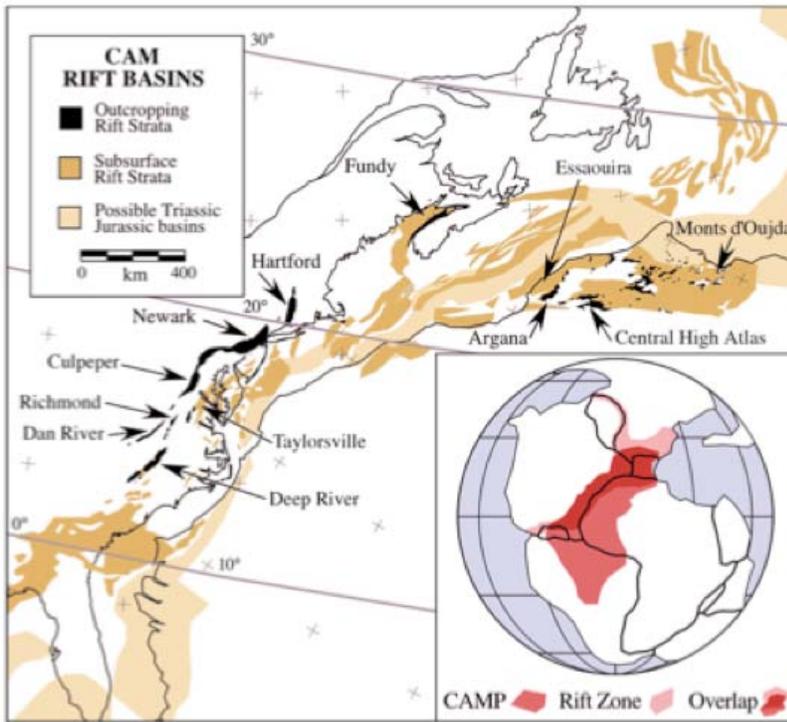


Figure 1. Central Atlantic Margin (CAM) rift basins in their paleogeographic position for the Carnian (~225 Ma). From Olsen and Et Touhami (2008).

Apart from the cyclicity in the CAM basin sequences, they also exhibit large-scale features such as lateral and vertical transitions in types of facies, which are produced primarily by tectonic processes. The current understanding is that the CAM basins are primarily half graben and half graben complexes and are bounded on one side by a major mostly normal fault system, towards which the basin strata tilt. Detailed studies of the Fundy Basin indicate that its Chignecto, Fundy and Minas subbasins are bound by normal faults to the north and the north-west (e.g. Withjack, et al., 1995). The complex tectonic history of the Fundy Basin started with an extensional episode in the Middle Triassic that reactivated the pre-existing Paleozoic compressional structures as boundary normal faults to the north. The second deformational episode resulted in shortening and compression in the NW-SE direction that inverted the subbasins.

As a result of the climatic cyclicity and the tectonic evolution of the region, Fundy and the other CAM basins show one or more tripartite sequences consisting of a “lower, relatively coarse-grained, fluvial facies with a rapid transition upwards into relatively fine-grained, deeper-water facies, followed by a slower transition upward into shallower-water and even coarse fluvial facies” (Olsen, 1997). Recent depositional models suggest that each tripartite sequence probably could represent a different tectonic episode i.e. a resurgence of extension or an acceleration of extension. Many such sequences are observed in the Atlantic conjugate margin basins, which suggest that their basin fill can be organized into discrete tectonostratigraphic sequences (Figure 2).

These tectonostratigraphic sequences can be used as a framework to understand the large-scale tectonic events that influenced the basin evolution and help discuss the rift sequences themselves. In the case of CAM basins, age is used as the criterion to group sequences in isolated basins into individual categories. A detailed description of the sequences can be found elsewhere (e.g. Olsen, 1997) so a shorter synopsis with their characteristics in the Fundy Basin is provided below:

Tectonostratigraphic Sequence I (early Middle Triassic)

The TS I sequence is made up of initial synrift sedimentary rocks that show strongly wedge-shaped sequences in small basins with beds fanning markedly towards faults. This sequence is present in the Fundy basin and forms two isolated set of outcrops along the Bay of Fundy (Figure 2). They are expressed as alternating sequences of mostly fluvial and aeolian dunes and minor red lacustrine mudstones, and comprise the the Honeycomb Pt. Formation in the Fundy Basin (Figure 3). These are the oldest dated strata in the Newark Supergroup.

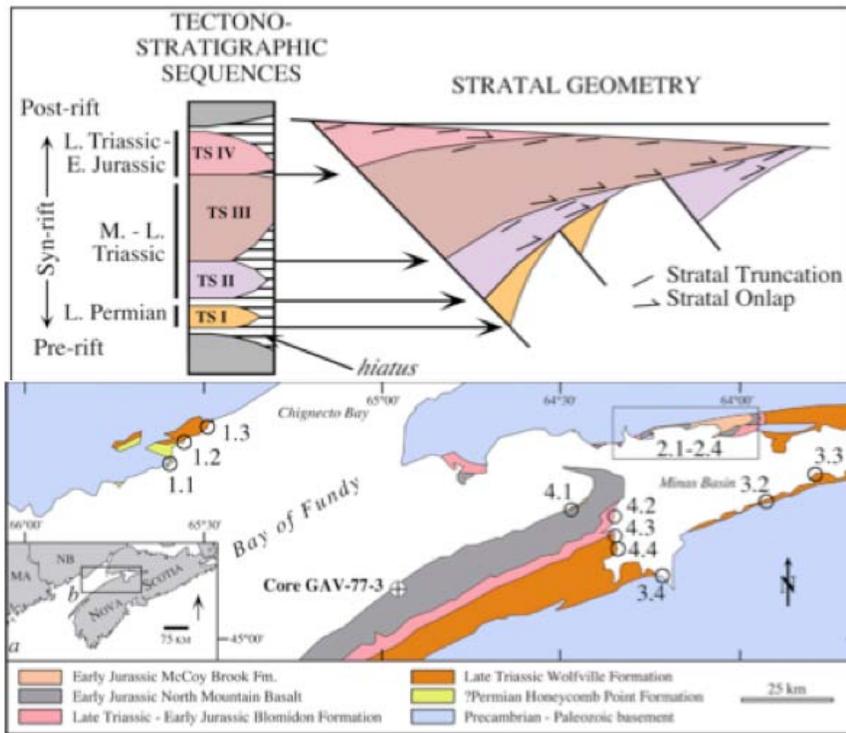


Figure 2. Tectonostratigraphic sequences of rift sequences in the CAM rift basins (top) and the location of the units in the Fundy Basin in map view (bottom). From Olsen and Et Touhami (2008).

Tectonostratigraphic Sequence II (Middle Triassic to early Late Triassic)

The TS II consists of early synrift sedimentary rocks that show wedge-shaped sequences in basins larger than those in TS I with lacustrine rocks that tend to be dominant in the basin's center. The TS II sequences exhibit evidence of deposition in more humid environments than those of TS I, as well as those of the younger sequences in the same basins. In the Fundy basin, the TS I is unconformably overlain by TS II, which otherwise rests on pre-rift basement. In the outcrops, TS II consists of fluvial strata of the Wolfville Formation in Nova Scotia. The eolian strata and minor interbedded lacustrine rocks usually grouped with the Wolfville Formation actually belong in TS III. The main mass of the Wolfville Formation consists of large (2–10 m) fining-upward cycles that were deposited by braided rivers, are extensively bioturbated, and have minor amounts of caliche. The upper Wolfville Formation consists of a lower fluvial sequence overlain by minor mudstones and often thick layers of eolian dune sands.

Tectonostratigraphic Sequence III (Late Triassic)

The middle synrift sedimentary rocks of TS III are geographically the most widespread units in the Fundy basin. This sequence thickens less towards the border faults in the north or northwest than any of the other tectonostratigraphic sequences. The thickness of these strata can exceed 6 kilometers. It is often comprised of a thick fluvial interval overlain by a thicker lacustrine sequence. In the Fundy basin, TS III is represented by the Blomidon Formation, overlying the upper Wolfville Formation by angular unconformity, at least on the local scale. The Blomidon Formation consists mostly of sand patch cycles, nodules and crystals of gypsum and occasional beds of eolian dune sand. The middle of the formation contains a sequence of interstratal karst deposits that could be dissolution sequences after halite and are associated with fish-bearing laminated mudstones. The upper few meters of the Blomidon Formation contain a sequence of thin gray and black plant-bearing mudstones. Together, these mudstones and the sand patch cycles make up a hierarchy of higher order cycles that are indicative of Milankovitch forcing (Olsen, 1997).

Tectonostratigraphic Sequence IV (Early Jurassic)

The late synrift TS IV begins with an extensive sequence of tholeiitic lava flows and interbedded sedimentary strata that overlie the TS III sequence. The sediments that are interbedded and overly the basalts typically have much higher sedimentation rates than underlying sequences. The TS IV is abundant in the in the Fundy basin and consists of the uppermost Blomidon formation, the North Mountain Basalt and the overlying McCoy Brook forma-

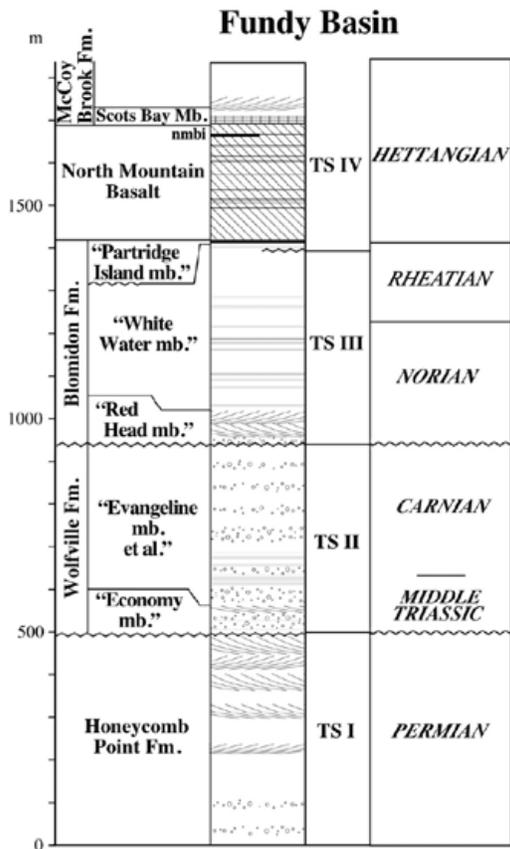


Figure 3. Stratigraphic column of the Fundy Basin with comparison to the tectonostratigraphic sequences of CAM rift basins. From Olsen and Et Touhami (2008).

tion. The uppermost Blomidon formation contains the Triassic-Jurassic boundary and has portions of dark shale beds. Although the radiometric dates from the lava flows are difficult to interpret due to excessive alteration, U-Pb dates from the North Mountain basalt of the Fundy basin show dates of about 202 Ma. The duration of the CAM basin igneous episode is around 600 ky and is therefore similar to that of other flood basalt provinces such as the Deccan and the Siberian traps. Overlying the Blomidon is a thin sequence of grey and white limestone and chert called the McCoy Brook formation. This formation is thought to have been deposited under generally more humid conditions than the similar-looking Blomidon formation based on the rarity of evaporites and sand patch mudstone and the abundance of bioturbation and fluvial-lacustrine deposits (Olsen et al., 2008).

The tectonostratigraphic sequences show simple changes in facies that are primarily controlled by tectonic events. However, some changes can be best explained by the northward drift of Pangea through stationary zonal climate belts (e.g. Robinson, 1973), which has implications for our understanding of the paleoclimate in the Mesozoic. Moreover, the sedimentary cyclicity produced by Milankovitch-type forcing allows calibration of basin processes and production of a Triassic-Early Jurassic time scale (Olsen, 1997). The Fundy basin and the other CAM basins therefore contain prehistoric records that may further our understanding of the paleoclimate and the paleotectonics in the Mesozoic.

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The Hartford Basin *Contributed by Jesse Farmer*

The Hartford Basin is one of several Mesozoic rift basins in eastern North America that formed during the early rifting of the supercontinent Pangaea. An east-tilted half-graben, the Hartford Basin is oriented north-south and comprised of fluvial and lacustrine sediments and basaltic lava flows dating from the late Triassic to the middle Jurassic (Figure 1) (Schliche and Olsen, 1989; Kent and Olsen, 2008).

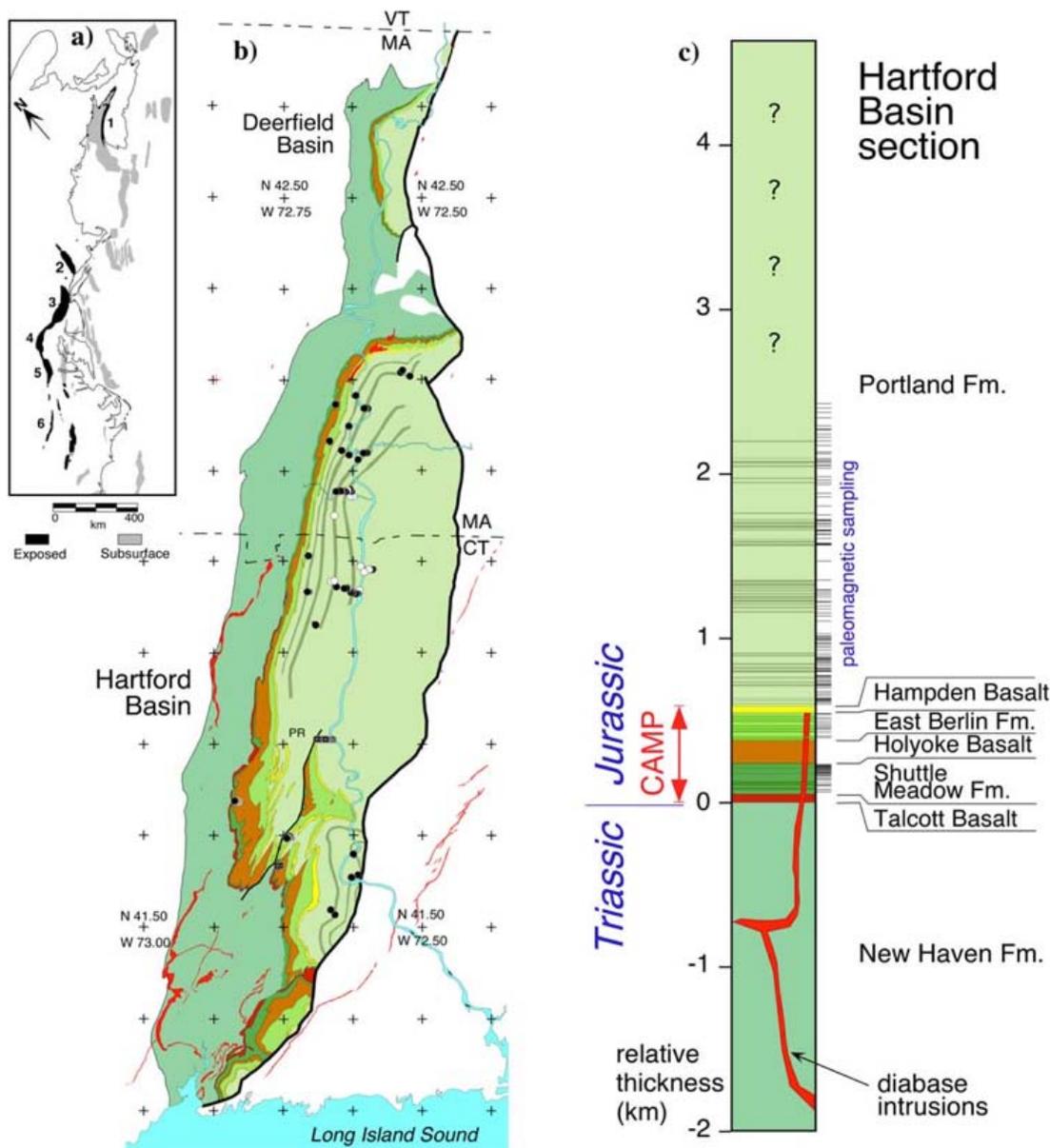


Figure 1. Geologic map (left) and stratigraphic column (right) of the Hartford Basin (from Kent and Olsen, 2008).

The stratigraphy of the Hartford Basin is given in Figure 1. The lowermost unit within the Hartford Basin is the ~2 km thick New Haven Formation, consisting of fluvial to lacustrine sandstones and mudstones with abundant soil carbonates and corresponding to tectonostratigraphic sequence III in the Fundy Basin. The uppermost New Haven formation and all overlying units corresponding to tectonostratigraphic sequence IV in the Fundy Basin. The uppermost New Haven formation contains evidence of the end-Triassic mass extinction and is overlain and intruded by flows of the Talcott Basalt. Pillows and sills of the Talcott Basalt, ~75 m thick, are overlain by the ~100 m thick lacustrine Shuttle Meadow Fm., which is in turn overlain by the ~200 m Holyoke Basalt flow. Above the Holyoke, the East Berlin Fm. consists of ~175 m of lacustrine sediments with pronounced cyclicity. A third basalt flow, the Hampden Basalt (~50 m thick), overlies the East Berlin Fm. Above the Hampden Basalt, the ~2 km thick Portland Fm consists of mudstones to coarse conglomerates consistent with fluvial to lacustrine depositional environments (Resor and deBoer, 2005).

East Berlin Formation

The East Berlin Formation is noted for both pronounced cyclicity in outcrop and spectacularly preserved dinosaur tracks. Sediments ranging from dark laminated, fossil-rich claystones to red, massively bedded siltstones and sandstones are interpreted as representing short-period, climatically driven fluctuations in lake levels (Figure 2). Fourier analysis of East Berlin stratigraphy shows dominant periodicities of 12.0 and 68.3 m (Olsen et al, 2005). The shorter of these periods corresponds to Van Houten cycles between darkly laminated layers; these have been associated to ~21 kyr precession cycles considering the control of precession on precipitation and lake levels in equatorial environments over the Pleistocene.

Abundant large theropod dinosaur tracks (*Eubrontes*) are excellently preserved in a mudstone unit of the upper East Berlin Formation (Figure 3). This facies corresponds to the uppermost Van Houten cycle and a regressive sequence from deeper lake conditions. These tracks have been postulated to represent the first appearance of large theropod dinosaurs (probably *Dilophosaurus*), which may be due to either an immigration event, or an abrupt evolutionary event resulting from the removal of top predators by the end Triassic extinction (Olsen et al., 2002).

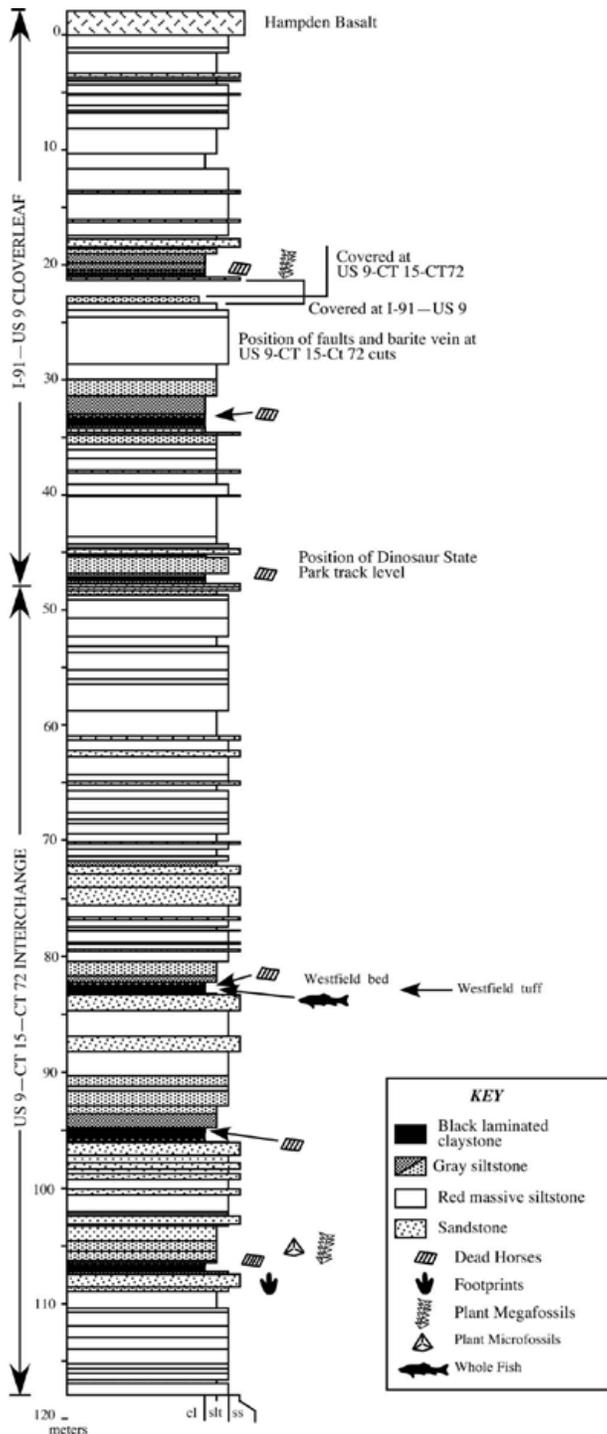


Figure 2. Stratigraphy of the East Berlin Formation, from Olsen et al. (2005).

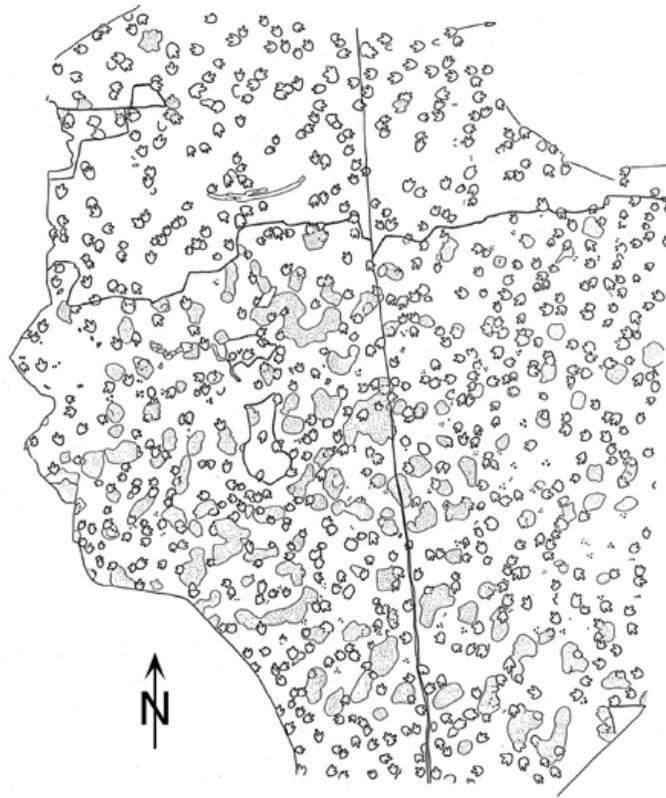


Figure 3. Dinosaur tracks at Dinosaur State Park outcrop (from Galton and Farlow, 2003).

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The Minas Fault Zone

Contributed by Jacob Eichenbaum-Pikser

The Minas Fault Zone (MFZ), comprising the Cobequid fault in the west and the Chedabucto fault in the east, marks the boundary between the Meguma and Avalon terranes in the Canadian Appalachians. It extends about 300 km on land, with direct exposures scattered from Cape Chignecto to the Chedabucto Bay, and associated deformation from Southern New Brunswick to Southern Cape Breton (Murphy et al., 2011). It terminates westward in the Gulf of Maine, and the eastward continuation has been proposed to extend all the way to the central Grand Banks, where it has been correlated with a prominent magnetic anomaly (Haworth and Lefort, 1979).

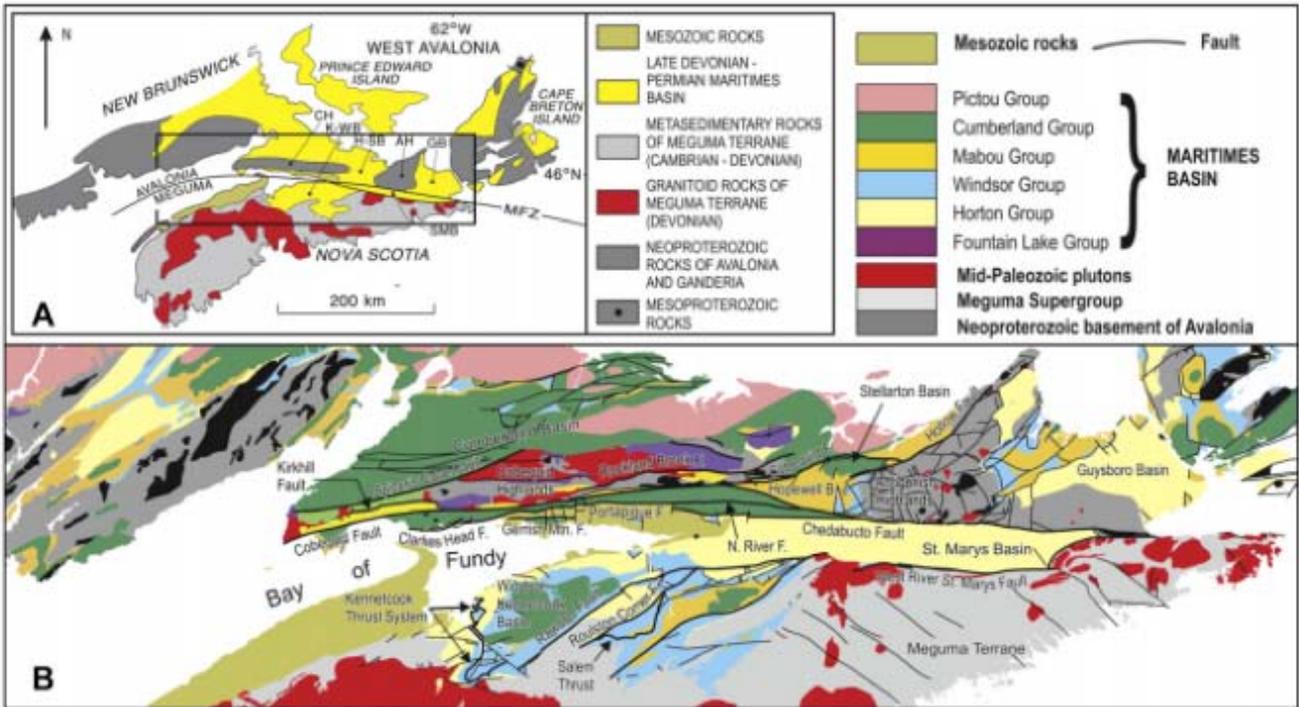
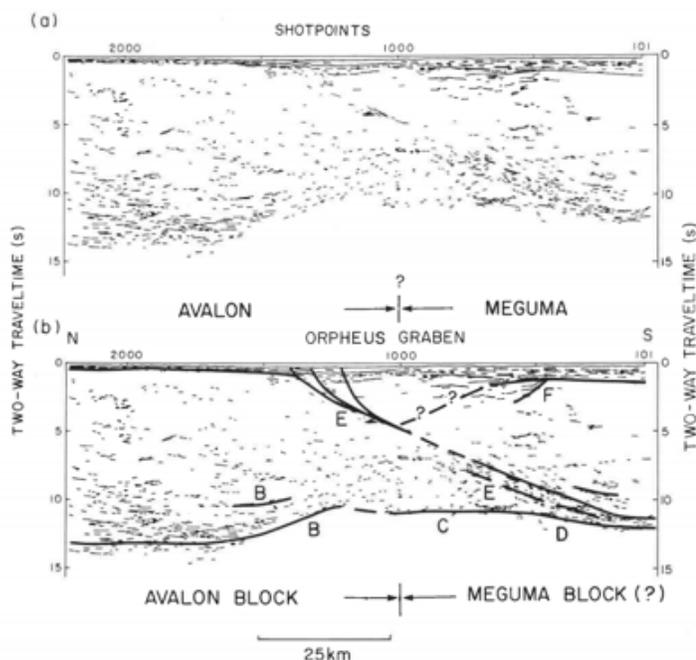


Fig. 1. A. General map of the Minas Fault Zone (MFZ), which separates Avalonia to the north from the Meguma terrane to the south. AH, Antigonish Highlands; CH, Cobequid Highlands (see Fig. 3); SMB, St. Marys basin (see Fig. 4); K-WB, Kennetcook-Windsor basin (see Fig. 5); GB, Guysborough basin (see Fig. 6); H-SB, Hopewell basin-Stellarton basin (see Waldron, 2004). B. Summary geological map of the geology in the vicinity of the Minas Fault Zone, (Nova Scotia geology based on Keppie, 2000; New Brunswick geology from New Brunswick Department of Natural Resources and Energy, 2000).

Figure 1. (above) General map of the Minas Fault Zone from Murphy et al. (2011).

Figure 2. (left) Seismic Profiles of the Minas Fault Zone from Marillier et al., 1989.



Seismic profiles (left) show the fault to be of crustal scale and listric geometry, with average dip of about 30°S to the Moho at a depth of 12 s (Marillier et al., 1989).

Deformation structures provide evidence of four discrete episodes of oblique dextral shear (both transtensional and transpressional) in the late Paleozoic. These events were accompanied by syntectonic magmatism and regional fluid flow, the latter leading to significant iron oxide copper gold (IOGC) ore deposits (Murphy et al., 2011).

Fig. 10. (a) Line drawing of profile 86-5B. (b) Interpreted line drawing of profile 86-5B. Major crustal thinning is observed beneath the Orpheus Graben. The fault bounding the basin to the north extends downward to the lower crust. The southern limit of the Orpheus Graben is indicated by a dashed line joining the maximum and the minimum basement depths across the basin according to the depth-to-basement map (Wade and MacLean, in press).

The most extensive outcrops of the MFZ are along the Cobequid Fault between Cape Chignecto and Parrsboro, where the fault extends 60 km along the north shore of the Minas basin before cutting inland. This scarp separates zones of relatively low strain (external zone) and high strain (internal zone), with outcrops of both accessible in the area (Waldron and White, 2005). The external zone outcrops at Cape Chignecto and is characterized by compressive structures. The lower degree of strain in this area has allowed for detailed interpretation, with four generations of folding identified. The internal zone outcrops in Greville Bay, and can be subdivided into three domains: the marginal domain contacts the external zone in a sharp transition to greater strain and localization, the core domain is a narrow (~100 m) zone characterized by yet higher strain and intense veining, and the shear-band domain is embedded discontinuously throughout the core domain, and represents the zone of highest strain on the fault. The principal unit in across these domains is Greville formation black phyllites (Waldron and White, 2005).

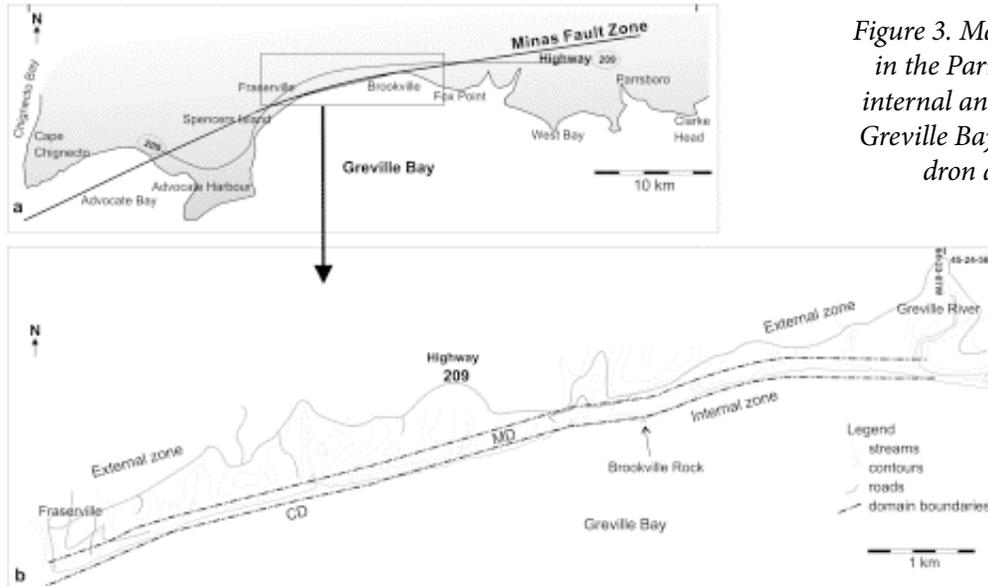


Figure 3. Map of Minas Fault Zone in the Parrsboro area (top) and internal and external zones along Greville Bay (bottom). From Waldron and White, 2005.

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Joggins Fossil Cliffs *Contributed by Anna Foster*

The name Joggins comes from the Mik'maq name Chegoggin or Chegoggins, meaning either 'Place of Weirs' or 'The Great Encampment'. Today, Joggins Fossil Cliffs is a UNESCO World Heritage Site, easily meeting the selection criteria for its uniqueness, scientific value, educational work, and preservation activities. Scientifically, Joggins is important for four main reasons: the fantastic individual fossils; the 'Grand Exposure' of the easily-accessible geologic record there, resulting from tilted Pennsylvanian strata exposed by tidal erosion (Figure 1); the in-situ nature of the fossil assemblages, allowing reconstruction of the paleo-environment; and the impact the location had on 'big ideas' of the late 1800s, particularly coal use in the Industrial Revolution and the theory of evolution. Joggins Fossil Cliffs is accessed through the museum, and both indoor and outdoor tours are available.

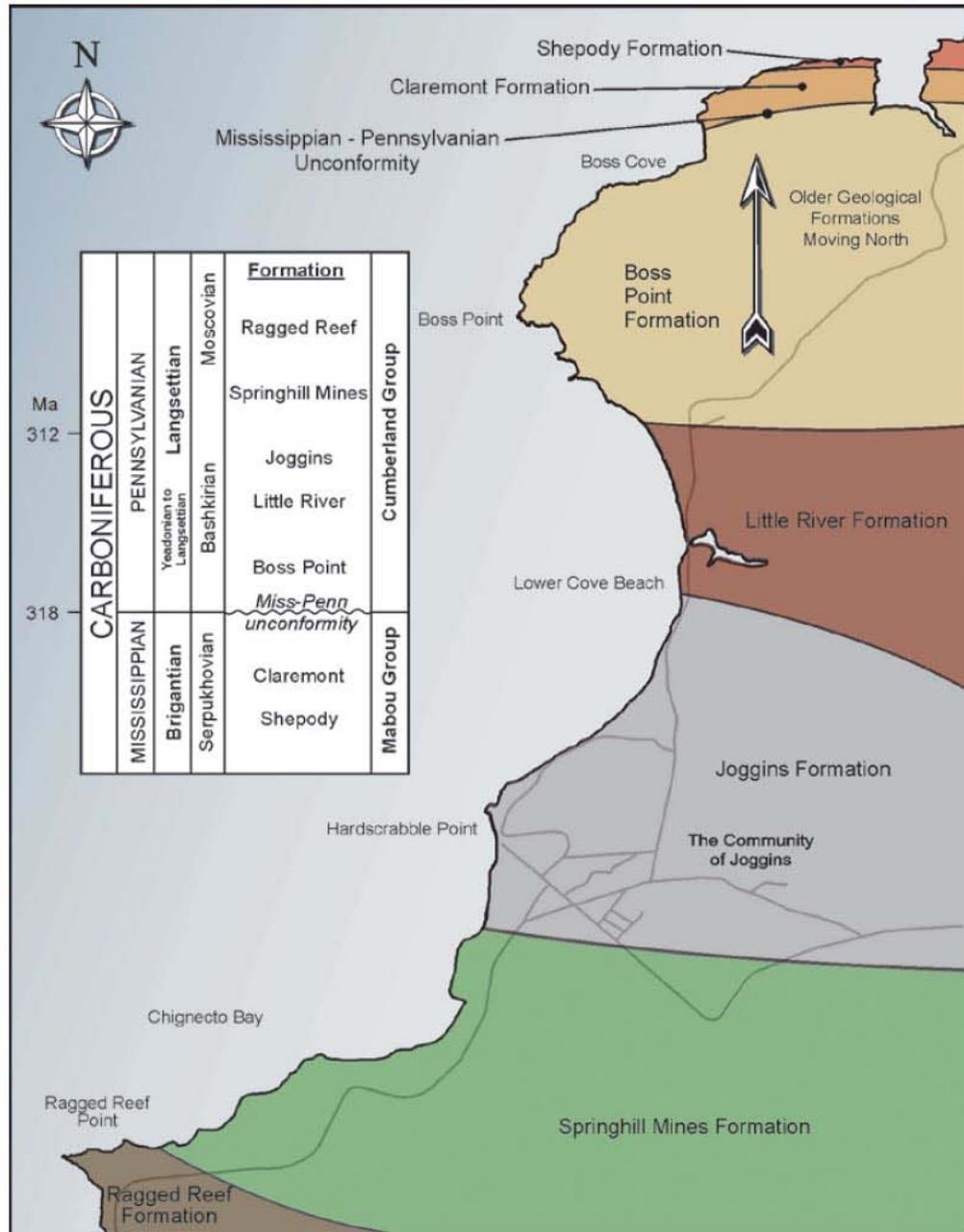


Figure 1. Geological Map of the Joggins Fossil Cliffs region. From Grey and Finkel (2011).

Historical Background

In the early 19th century, the Industrial Revolution made coal into a valuable commodity, albeit one of unknown origins. The debate of the time was whether the formation of coal occurred from in-situ plant matter or drifted materials. The first published geological account of Joggins was by mining engineer Richard Brown in 1829, and documented the upright fossil trees in the cliffs. Similar fossils had been discovered in Britain and Europe; Sir Charles Lyell examined the Joggins fossils in 1842, linking the upright trees to the coal seam layers. This, combined with observations of rooted underclays in coal seams in Wales and Pennsylvania by Sir William Logan settled the debate, with coal formation accepted to be in situ.

With the geologic importance of Joggins now established, and a new job as the Nova Scotia provincial geologist, Logan made the first log of the Joggins section, totaling over 4 kilometers, in only 5 days. Lyell returned to Joggins with Sir William Dawson in 1852, and discovered fossils of early amphibians (*Dendrerpeton*) and land snails (*Dendropupa*) inside a fossilized tree. This revolutionary discovery proved that the Joggins fossils were not only well-preserved individually, but also in relation to the other fossils. This spurred the creation of a detailed log of the Joggins section, begun by Lyell and Dawson and completed by Dawson, who spent the next 50 years largely studying the Joggins fossils. In 1859, Dawson discovered the earliest known reptile within a fossil tree and named it in honor of Lyell—*Hylonomus lyelli*. Dawson was assisted in his research by mining companies working in the area, who brought many specimens to his attention. For further historical details, see Falcon-Lang (2006) and references therein.

Formation and Preservation

As mentioned, the fossils at Joggins are preserved in parautochthonous assemblages, meaning they are in context. Additionally, the entire section was deposited in less than one million years, and collector curves based on the documented 150 years of observation suggest that the Joggins fossil record is relatively complete (Falcon-Lang et al., 2006). Thus, by analyzing taxa within a single facies, the paleo-ecosystem can be reconstructed.

The results of such studies are three different environments that interchanged with one another in response to base-level fluctuations forced by tectonism and glacioeustasy: brackish seas, poorly-drained coastal plains, and well-drained alluvial plains (Falcon-Lang et al., 2006). Within the Joggins section, 14 ‘rhythms’ of these interchanges are observed, cycling through:

- Retrograding poorly-drained coastal plain, characterized by the thickest coal seams. This environment was populated by foraminifera, molluscs, annelids, arthropods, fishes, and aquatic tetrapods.
- Open-water brackish seas, today seen as limestone overlain by sandstone/mudstone, with similar species to those observed in the retrograding coastal plain.
- Prograding poorly-drained coastal plain, covered in rainforest and the source of the upright tree fossils, Lycopside, as well as the flora calamiteans, ferns, pteridosperms, and cordaitaleans, and the fauna molluscs, annelids, arthropods, and tetrapods, including *Hylonomus lyelli*.
- Well-drained alluvial plains covered in fire-prone cordaitalean scrub. The tropical forests produced a lot of oxygen, which was important for building up the present-day atmosphere, but also made fires from lightning strikes common. This resulted in the selective preservation of fossils in waterholes from this environment, mainly molluscs, arthropods, and tetrapods.

As can be observed by the height of the fossil tree trunks (Figure 2), the trees were buried multiple meters deep before they began to rot. This requires extremely high accumulation rates, with estimates of 40 years to accumulate 1 meter of sediment. At the time of deposition, Joggins was located in the Cumberland sub-basin connected to the ocean, with the Caledonian Mountains to the west in present-day New Brunswick and the Cobequid Mountains to the east in present-day Nova Scotia, both ranges higher than the Rocky Mountains. The basin was underlain by the Mississippian Windsor Group limestones and evaporites, described as a lake bed with repeated marine incursions, notably including a 250 m thick salt deposit. The withdrawal of this salt at depth is primarily

responsible for the extraordinarily high accumulation rates of sediments shed off the surrounding mountains, with some contribution from subsidence due to motion on basin-bounding faults (Waldron & Rygel, 2005).

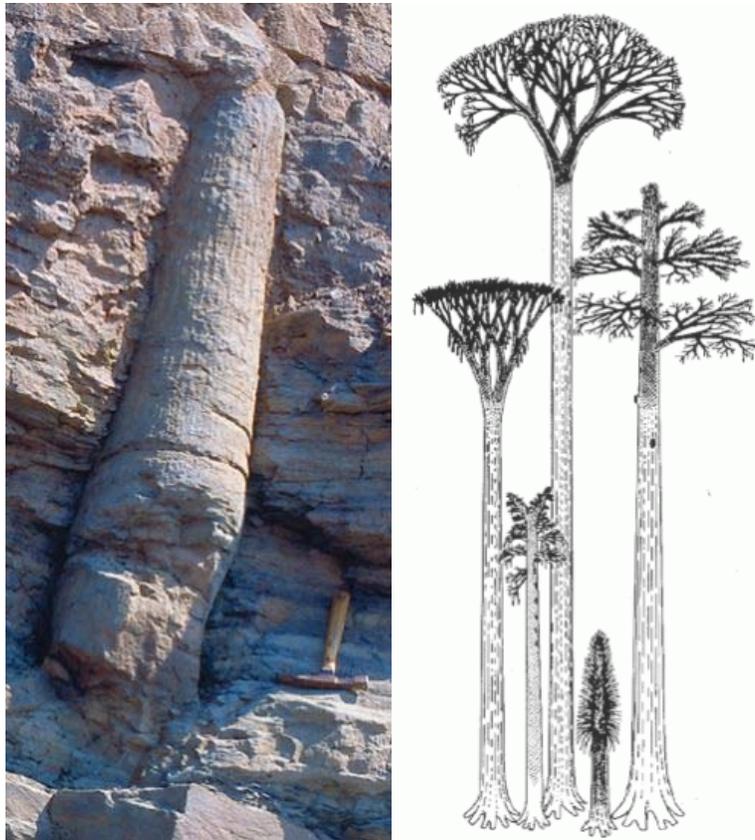


Figure 2. *Lycopodioid* fossil tree preserved in the Joggins Cliffs (left); reconstructed *Lycopodioid* trees (right). From <http://jogginsfossilcliffs.net/cliffs/biodiversity/>

The most recent evolution of Joggins occurred approximately 10,000 years ago, when glaciation during the last Ice age compressed the area by 25 meters, then covered it with glacial till as the ice withdrew (JFC Tour Guide, personal comm., 2012). At approximately the same time, the bay reached Joggins, beginning the period of an extremely high rate of erosion that can be observed today, estimated by Dawson to be 19 cm per year (Dawson, 1882).

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The Arisaig Group

Contributed by Jesse Farmer

The Arisaig Group comprises 1500 to 1750 m of strata exposed along the shores of Northumberland Strait in northeastern Nova Scotia, dating from the latest Ordovician to the early Devonian (Figure 1). Separated into eight formations, the Arisaig Group is generally characterized by mudstone to siltstone facies representative of sediment deposition in a nearshore basin. This basin was part of the microcontinent of Avalonia, which docked with Laurentia during the mid to late Silurian. The overall subsidence history of the basin is characterized by two intervals of tectonically driven subsidence, the first of which may reflect the oblique collision between Avalonia and Laurentia, and the second of which may reflect collision of Avalonia with the Meguma terrane to the south (Figure 2). Facies changes superimposed onto these two subsidence events suggest a record of global eustasy and regional sea level changes associated with basin infilling and tectonic loading. As a result, sediments from the Arisaig Group are an important constraint for global eustatic sea level records during the Ordovician to Silurian interval.

Figure 1. Stratigraphy of the Arisaig Group (from Melchin and MacRae, 2005). Labeled stops are those units observed on this field trip.

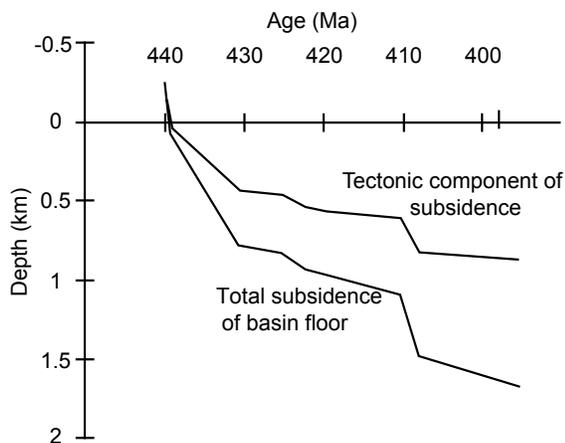
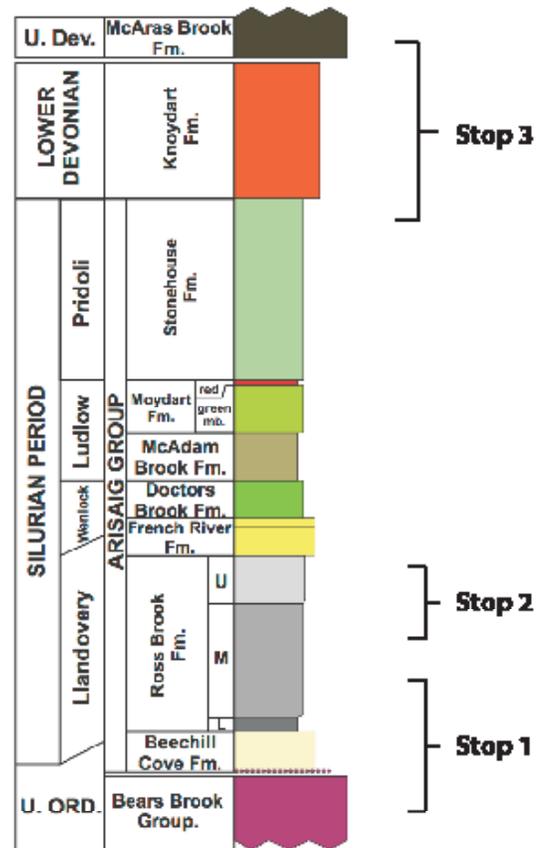


Figure 2. Subsidence history of the Arisaig Basin (from Waldron et al, 1996)

Chronology for the units of the Arisaig Group is established by brachiopod and graptolite biostratigraphy and palynological data, supplemented by chitinozoan assemblages in the lower units, and bryozoan, gastropod, trilobite and crinoid assemblages in the upper units. Radiometric ages are available from basal volcanics of the Bears Brook Group, constraining the lowest sediments of the Arisaig Group to later than middle-upper Ordovician, and from the upper McAras Brook volcanics, which constrain the latest Arisaig sediments to the Upper Devonian. Support for the nearshore depositional environmental interpretation comes from common, coarse fossiliferous layers throughout the Arisaig Group, which are interpreted to represent storm layers.

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Cape Breton Highlands (modified from Parks Canada-Cape Breton National Park Webpage)
 Contributed by Rafael Almeida

To understand the geology of Cape Breton Island (Fig. 1), we must first look at how the geological landscapes (terranes) that form the island today came together. Plate tectonics has played a major role in forming the island of Cape Breton.



Figure 1. NASA Landsat image of Cape Breton Island

Three terranes make up Cape Breton Island (Fig. 2). The Blair River inlier on the northwest tip is composed of the oldest rocks known in the Maritime Provinces. The Bras d'Or terrane which makes up most of the northern half of Cape Breton Island is a series of sedimentary and volcanic rocks that began forming off the northwest coast of the continent we now call South America. The Avalon terrane, which makes up the southern half of Cape Breton Island, is volcanic rock that first began forming on the northwest coast of what is now Africa.

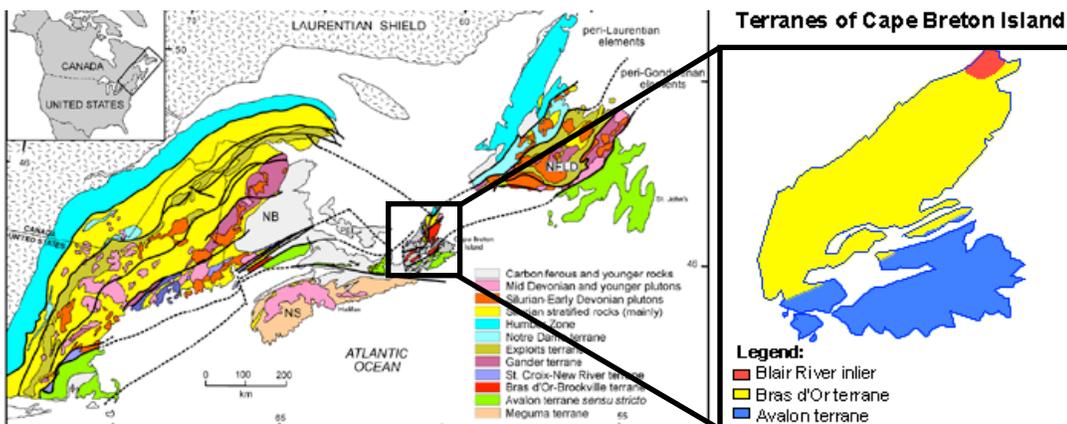


Figure 2. On the left, simplified geological map of the Maritime Provinces showing terranes in the northern Appalachian orogen (from Barr et al., 2005) and on the right, further zoomed in on Cape Breton Island (from Parks Canada 2006)

1,500 to 750 Million Years Ago - The oldest rocks in the Maritime Provinces form the Blair River inlier which is found in the northwestern corner of Cape Breton Island. They can be seen on North Mountain as you drive up from Pleasant Bay. These rocks were formed 1.5 to 1 billion years ago during the collision of continental plates that resulted in the supercontinent Rodinia. They are probably a continuation of the Canadian Shield which is found in parts of Labrador, Québec, Ontario and Nunavut. As such, they are likely part of the core of the North American continent (Fig. 3).

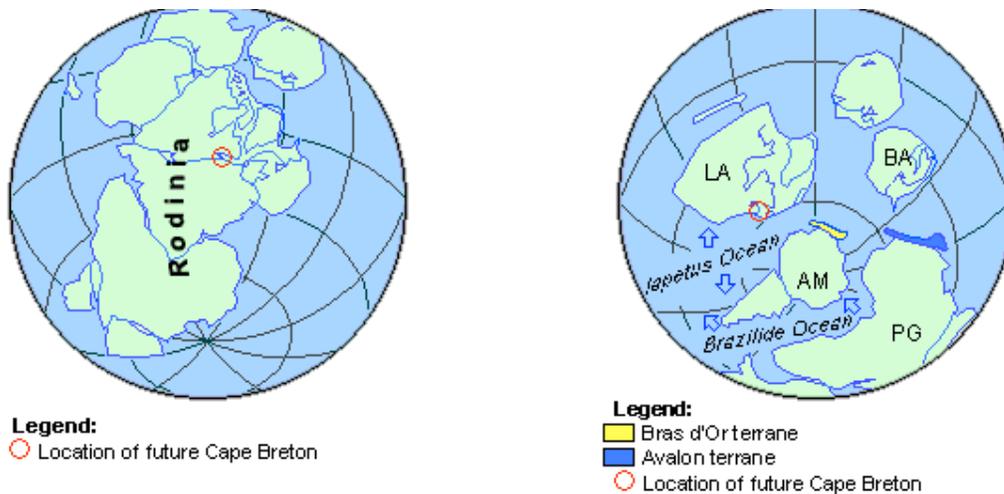


Figure 3. Map on the left shows paleogeography at 850 Ma (Neoproterozoic) after the assembly of Rodinia. Map on right shows paleogeography at 540 Ma (Cambrian) after the breakup of Rodinia.

750 to 450 Million Years Ago – During this time Rodinia broke up into smaller continents which geologists call Laurentia (LA), Amazonia (AM), Protogondwana (PG) and Baltica (BA). These continents were separated by growing oceans. The Brazilide Ocean, found between Amazonia and Protogondwana, was the earliest to form. About 150 million years later, the Iapetus Ocean began to open up between Laurentia and Amazonia. As the early continents migrated, island arcs formed off the coast of Protogondwana, giving rise to the volcanic rocks of the Avalon terrane. Sedimentary rocks laid down in the seas off Amazonia that were later intruded with igneous rocks formed the Bras d’Or terrane. The Bras d’Or terrane and the Avalon terrane were driven together as the Brazilide Ocean closed about 580 to 450 million years ago to form Gondwana (Fig. 3, 4).

450 to 360 Million Years Ago - Through this time period, continents collided again. Baltica bumped into Laurentia to form a new continent called Euramerica. The Bras d’Or and Avalon terranes were dragged along by Baltica to become part of Euramerica too. The Blair River inlier, formed a billion years earlier and lying on the edge of the Laurentian Plate, was finally sandwiched together with these other two terranes to form the core of what we now know as Cape Breton Island (Fig. 4).

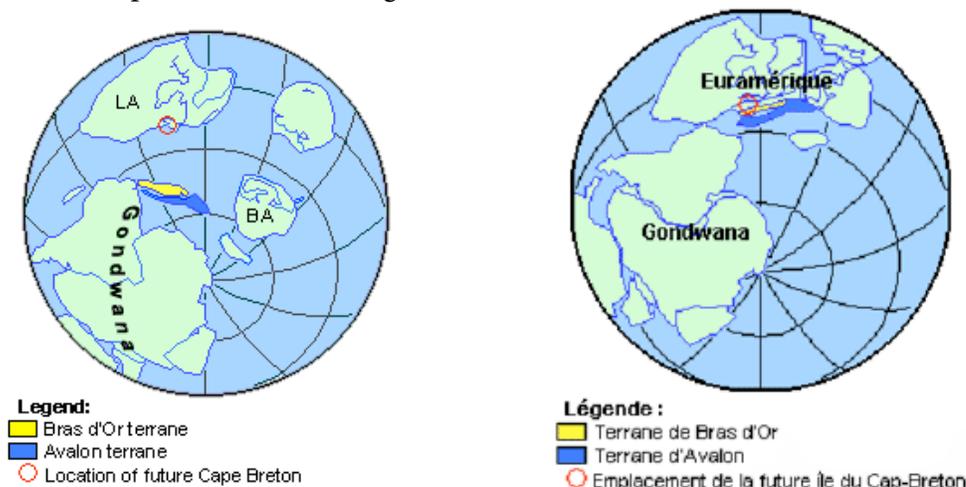


Figure 4. Map on the left shows paleogeography at 475 Ma (Ordovician) after the closure of the Brazilide Ocean. Map on right shows paleogeography at 435(Silurian) Ma after formation of Euramerica.

360 to 250 Million Years Ago - Starting about 360 million years ago, the ocean separating Euramerica from Gondwana slowly closed, bringing these two continents together near the equator. The collision pushed up and folded the seabed along what is now the east coast of North America, creating the Appalachian mountain chain. Faulting associated with this mountain building is now exposed in many of the canyons that characterize the modern Cape Breton Plateau. Between the up-folded seabed, deep basins were created. Much of the present-day Maritime provinces were located in such a basin.

From 340 to 325 million years ago sea level rose and fell, flooding this basin as the ocean closed and giving rise to alternating layers of sedimentary rocks formed from the salts left behind by evaporating water, such as gypsum, and red sandstone, conglomerate and shale. Several gypsum cliffs, which can be seen around Dingwall north of the park, are relics of this time.

By 325 million years ago, Euramerica and Gondwana were united as the supercontinent Pangea, with the Appalachians running down the seam which joined them. Most of the northern Maritimes except for northern Cape Breton Island was lowland covered by a tropical forest swamp which later became coal fields such as those found throughout Cape Breton Island.

About 290 million years ago the climate changed dramatically, putting an end to the tropical forests and replacing them with a brick red desert. There are no rocks of this age in present-day northern Cape Breton Island because the desert didn't extend up into the Appalachian Mountains, but the famous red soil of Prince Edward Island comes from rocks of this period.

250 Million Years Ago to Present - By around 250 million years ago the Maritime provinces were joined to what are now Morocco and Spain, at the centre of the Pangean supercontinent. As Pangea started to pull apart, the Maritimes were found on the west side of the slowly spreading Atlantic Ocean, on the edge of a new continent: North America. Erosion, deposition and formation of sedimentary rocks continued along the trailing edge of the North American continent. However, the sedimentary rocks dating from this era all formed to the east on what is now the Scotian Shelf and the Grand Banks, and are submerged.

While dinosaurs probably did walk the hills of northern Cape Breton Island, there is no trace of them because the hills have long since eroded, wearing away any trace of dinosaur bones and footprints. This erosion on the edge of the North American continent cut the whole of Atlantic Canada down to a broad, flat valley. Part of this ancient valley plain is now the flat top of the Cape Breton Plateau.

The Last 2 Million Years - The Atlantic Ocean continued widening and the North American continent drifted to its present location. It continues to drift an average of a few centimeters a year toward the northwest. Repeated glacial advances - ice ages - through the last two million years and ending about 10,000 years ago caused modifications to the already eroded landscape, carving out U-shaped valleys, shallow lakes, deep scratches called striations, and leaving behind extensive deposits of mixed mud, silt, sand and gravel called glacial till.

The Bras d'Or Lakes were carved out of the bedrock by glaciers during the ice ages. Many other geological and topographical features of Cape Breton Island also owe their form to the numerous glacial advances that occurred over tens of thousands of years.

The glaciers of northern Cape Breton Island were not always part of the continental ice sheet. At times there was only a small ice cap on the highlands, with the rest of northern Cape Breton ice-free.

The dominant feature of northern Cape Breton is the Cape Breton Plateau, averaging 350 metres at its edges but rising to more than 500 metres at its centre. The Cape Breton Plateau is part of the worn down Appalachian

mountain chain which stretches from Georgia to Newfoundland. Extending over 70% of the Cape Breton Highlands National Park, the plateau appears flat-topped but actually consists of broad, gently rolling hills, deeply cut by steep-walled river canyons and broad valleys. On the west coast the plateau meets the Gulf of St. Lawrence in steep cliffs. On the east, the highlands border the Atlantic Ocean in a more gently sloping coastal plain with low headlands and a few long sandy beaches (Fig. 5).



Figure 5. View of the Cabot Trail on the west coast of Cape Breton. Picture on the left is looking south (<http://homepage.usask.ca/~mmw819/Cabot%20Trail.html>). Note the flat topped hills of the Cape Breton Plateau. Picture on the right is looking north.

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Regional stratigraphy and tectonic setting

The Gulf of St. Lawrence Basin of Atlantic Canada is attributed to extensional collapse immediately following the mid-Devonian Acadian Orogeny (Stockmal et al. 1990), with subsidence continuing throughout the late Devonian to early Permian (Howie & Barss 1975). During the Late Carboniferous, eastern Canada was affected by dextral strike-slip faulting which reactivated pre-existing structures including the Belle Isle Fault, Cobequid–Chedabucto Fault and the Hollow Fault and also resulted in the formation of several small basins which underwent rapid tectonic subsidence (Fig. 1a, b, c). This deformation, which correlates with the onset of Alleghanian deformation in the USA (Gibling 1995), formed the St. Georges Bay Basin, which extends NE from St. Georges Bay into western Cape Breton Island along the trend of the Hollow Fault (Fig. 1a, b). The St. Georges Bay Basin is considered an extensional basin related to Late Carboniferous movement along the Hollow Fault, which lies immediately to the north of a long salt wall, and dips toward the south with a throw of c. 1650 m (Fig. 1b, c). Salt structures are especially prominent under the Gulf of St Lawrence, along the Hollow Fault and on-shore in Cape Breton (Boehner 1986; Howie 1988; Langdon & Hall 1994) (Fig. 1a, b, c). The tectonostratigraphic framework is related to the combined effects of regional subsidence of the Gulf of St Lawrence Basin (Late Devonian–Stephanian) and enhanced Late Carboniferous (tectonic) subsidence during the Namurian to Westphalian D/Stephanian.

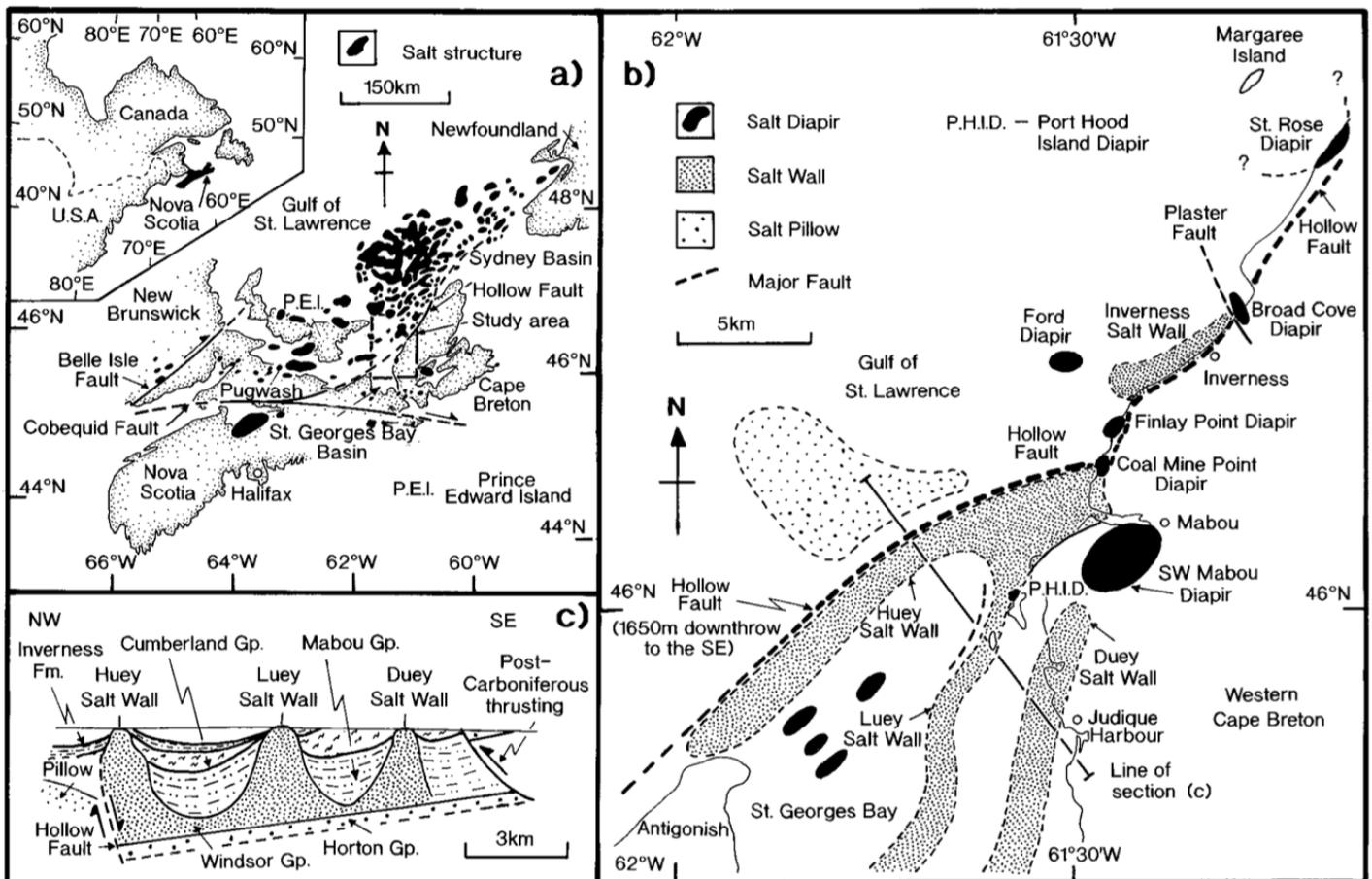


Figure 1. (a) Tectonic map of Nova Scotia showing major faults, the location of the study area (box) and salt structures in the Gulf of St Lawrence (modified from Boehner 1986). The inset map shows the location of Nova Scotia. (b) Map of St. Georges Bay and western Cape Breton diapir field, showing major salt walls, diapiric structures and on-shore localities referred to in the text. (c) Section through the major salt walls based on offshore seismic and onshore outcrop data. See (b) for the position of the section. (From Alsop et al. 2000)

The stratigraphic column for the area can be seen in Fig. 2. The diapir source unit is the (c. 1000 m thick) Windsor Gp of Visean age, described in detail most recently by Giles et al. (1997a, b). The group commences with a laminated limestone (the Macumber Fm), overlain by up to 400 m of halite and 300 m of anhydrite (Fig. 2). The upper part of the Windsor Gp comprises regionally extensive thin, carbonate beds, intercalated with red and grey mudstones and sandstones, as well as minor evaporites. The conformably overlying Mabou Gp (<1000 m thick) of latest Visean to Namurian age, comprises a lower Hastings Fm which is dominantly grey shales with minor limestone, siltstone and sandstone, and an overlying Pomquet Fm consisting of red shale, sandstone, with minor conglomerates. The Cumberland Gp consists of the Port Hood Fm (? Namurian to Westphalian A), which rests disconformably or unconformably on the Mabou Gp and is up to 3000 m thick (Fig. 2). The overlying Henry Island Fm (Westphalian B/C) is c. 400 m thick, and the Inverness Fm (Westphalian C/D to Stephanian), rests unconformably on older units and may be as much as 1000 m thick. The Broad Cove Fm conformably overlies the Inverness Fm and has a thickness of 200 m. The group is formed by sandstones, mudstones, shales and coals. Analysis of palaeocurrent data and the presence of conglomerates and unconformities in the vicinity of salt structures suggests that Upper Carboniferous sedimentation was affected locally by diapir growth (Haites 1952; Boehner 1986; Howie 1988). Analysis of seismic profiles that terminate in the near shore area has allowed on-shore geology and deformation zones to be placed in context with the parent diapir (Brown, 1998) (Fig. 3).

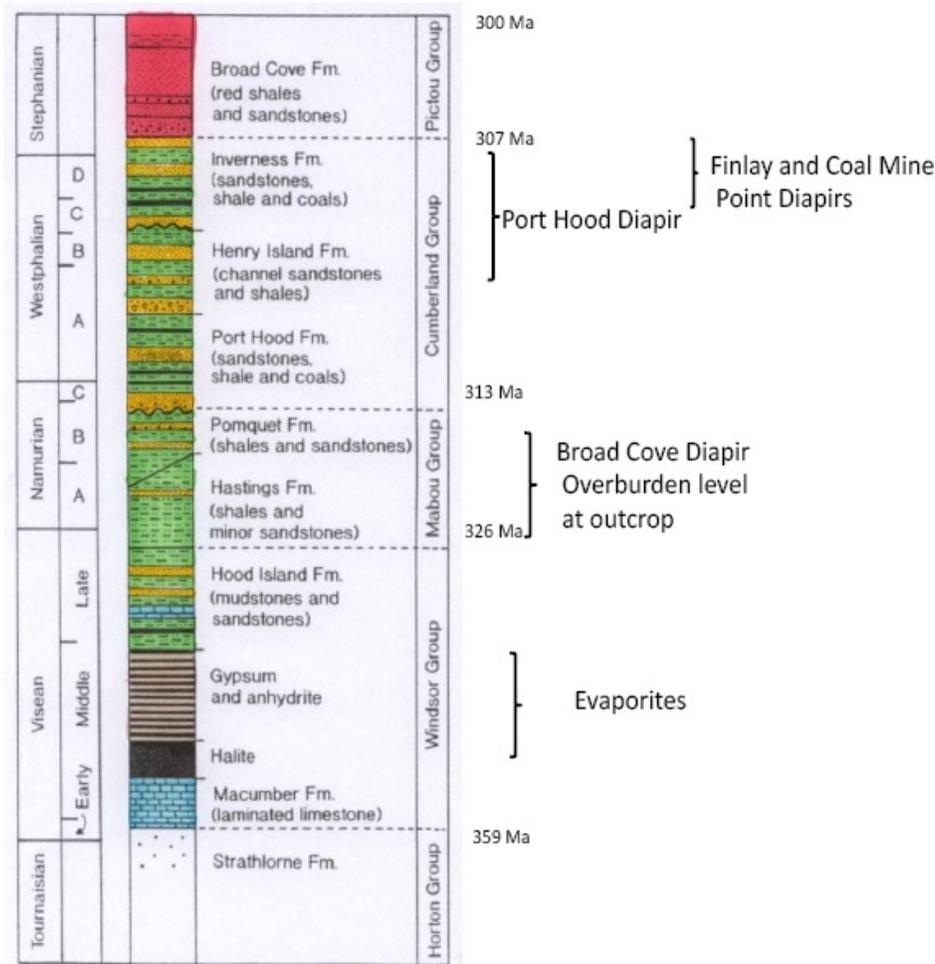


Figure 2. Major stratigraphic divisions of the Carboniferous rocks of the study area in western Cape Breton (based on Giles et al. 1997a, b). (From Davison, 2005).

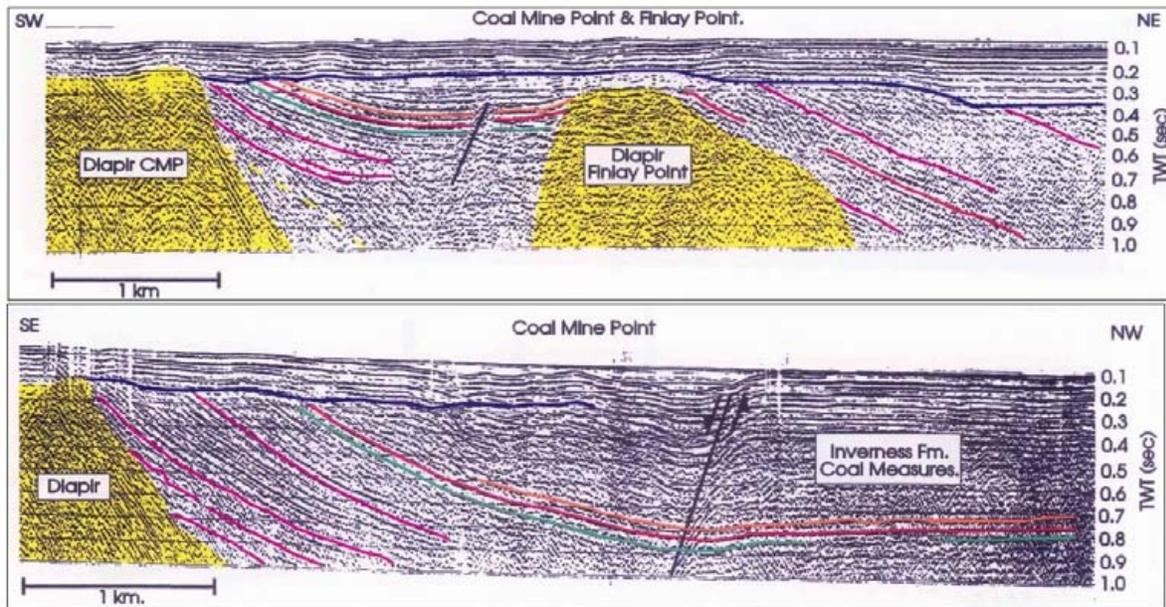


Figure 3. Section across Coal Mine and Finlay Point Diapirs. Upwarped beds above salt are parallel bedded and folded during later compression after diapir intrusion. (From Brown 1998).

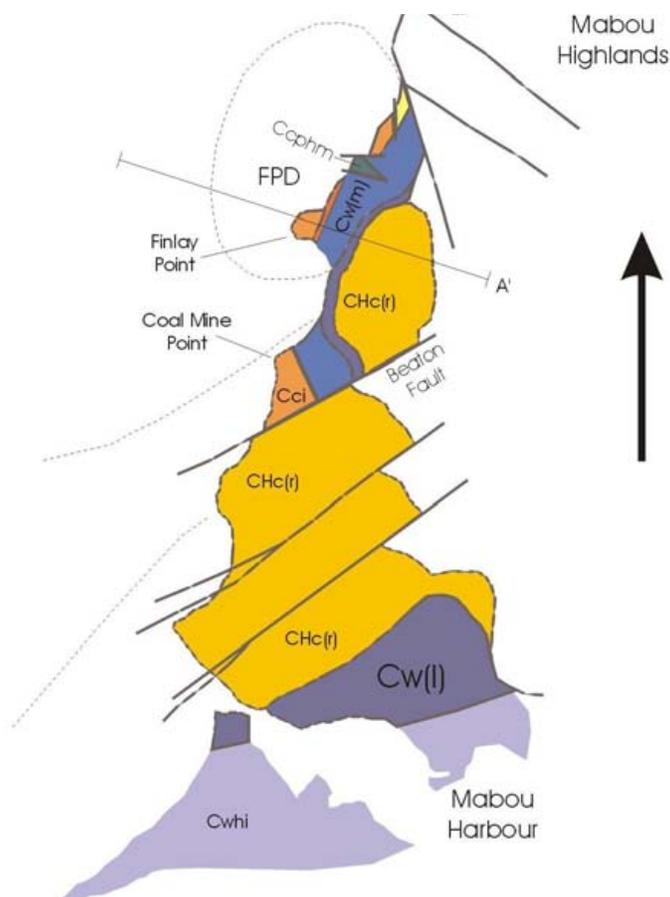


Figure 4. Geological Map of the Mabou Mines Area, Western Cape Breton. (From Davison, 2005)

The exposed sedimentary carapaces in western Cape Breton are characterized by intensely folded and faulted upper Windsor shales, thin sandstones and limestones, with irregular masses of gypsum (Brown 1998). In all the documented cases, upper Windsor strata are visible at outcrop in the structural carapaces, with evaporites documented from drilling or inferred to be present on stratigraphic grounds. The widespread development of cap rocks, which represent insoluble residues concentrated during dissolution, above diapirs both in Nova Scotia and in evaporite provinces elsewhere suggests an appreciable removal of halite.

The main diapiric growth phase appears to have occurred in the Westphalian–Stephanian or later in response to late/post Carboniferous inversion. Diapirs located within St Georges Bay show evidence of relatively early active (upbuilding) growth during the Namurian that was followed by a prolonged period of downbuilding diapir growth (passive growth, when salt layer feeding diaper subsides) through to the Stephanian (Brown 1998). Seismic data indicates that the diapirs are typically asymmetric structures between 0.9 km and 1.5 km high, e.g. Ford, SW Mabou, St Rose and Broad Cove diapirs (Fig. 1b). The

geometries described are not salt-cored anticlines as active penetration by the salt is clearly observed as is constrictional flow of salt up diapiric necks. Both the upbuilding and downbuilding mechanisms generate an upward movement of salt relative to the overburden resulting in the development of drag zones. Fig. 4 represents a simplified geologic map with the two diapirs observed in the trip, the Finlay Point and Coal Mine point diapirs.

Finlay Point Diapir

The Finlay Point Diapir is a markedly asymmetric diapir with a steep normal fault on its NW margin and the Hollow Fault forming the SE margin (Fig. 5a, b). Correlation of the onshore section to near-shore seismic data suggests that the mapped drag zone forms on the gently dipping flank of the diapir. The Inverness Fm has been deformed around the NE diapiric contact, but with no actual diapiric penetration suggesting that diapiric growth did not commence until after deposition of the Inverness Fm (Westphalian C–Stephanian). The 25 m thick Finlay Point Conglomerates adjacent to the diapir are overlain by weaker sandstones, shales and coals (Inverness Fm) typically intercalated on a meter scale. The onshore exposure of the Finlay Point Diapir consists of a massive white gypsum mylonite which can be traced inland into karst topography (46°08'07"N 61°2'46"W) (Figs 5c, 6). The margin of the diapir is marked by highly strained NE–SW-trending gypsum with a foliation which dips steeply to sub-vertically towards the NW, and contains a gypsum-elongation lineation which plunges steeply towards the WSW. There is a progressive steepening of the gypsum foliation into sub-vertical attitudes towards the diapiric contact which is orientated 170°/85°W and marked by a 1 m wide zone of brecciation related to normal faulting (Fig. 5c). The Finlay Point Diapir drag zone is exposed in a continuous 300 m section, and displays constant ENE–WSW strikes together with a pronounced increase in bedding dips across a NNE-trending fault, which separates the overall drag profile in to inner and outer zones (Fig. 5a, b, c). The inner drag zone is restricted to within approximately 40 m of the diapiric contact, with dips within thin shales (<1 m) and red conglomerate beds (2–3 m) not exceeding 40° towards the NNW, although they are locally steeper within faulted zones (Fig. 5c).

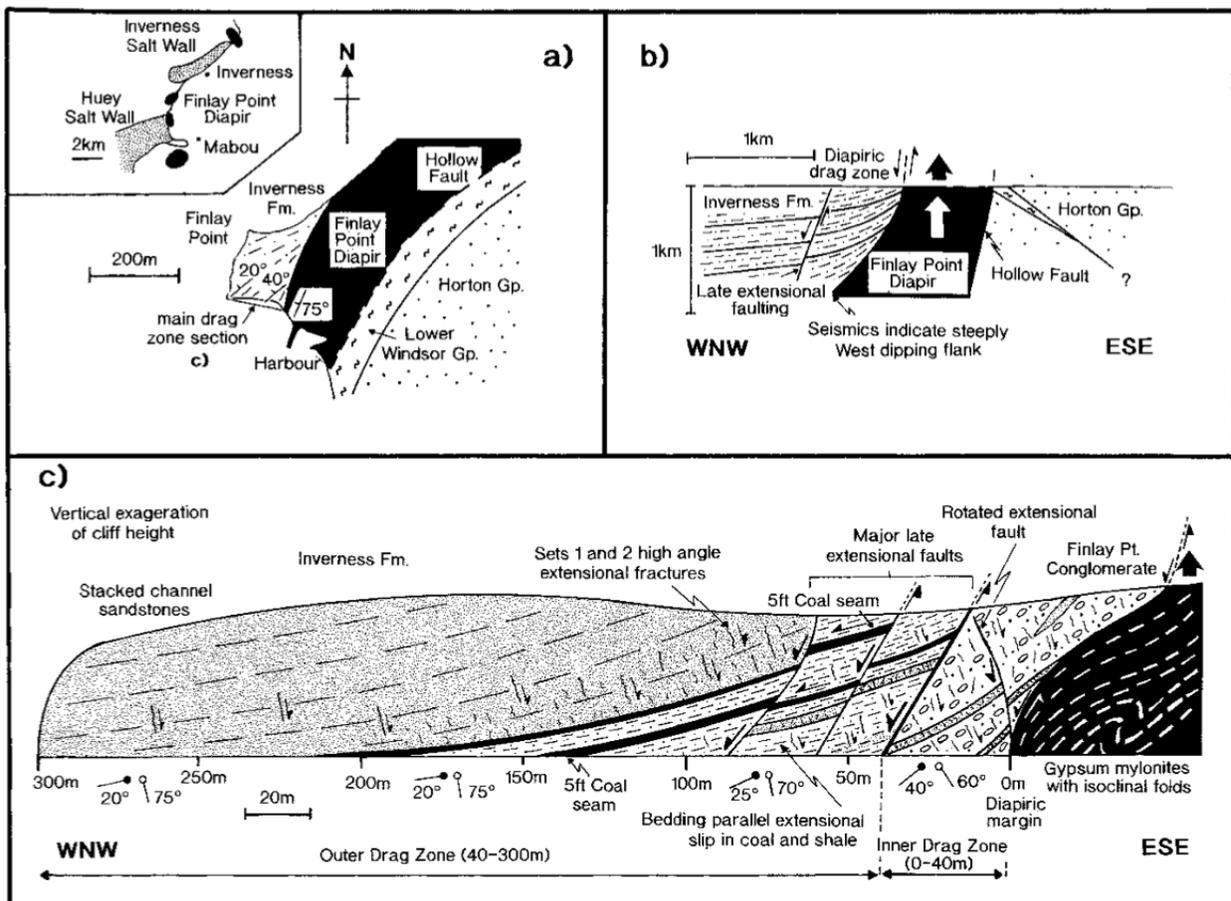


Figure 5. (a) Simplified geological map of the Finlay Point Diapir and adjacent areas (modified from Giles et al. 1997a, b). The map highlights the position of the main coastal drag zone section shown in (c), whilst the inset map shows the general location of the diapir (refer to Fig. 1b). (b) Schematic WNW–ESE section illustrating the overall geometry of the Finlay Point Diapir based on outcrop studies and offshore seismic data. The relative location of the detailed drag zone section (c) is noted. (c) WNW–ESE section through the Finlay Point Diapir drag zone showing rotation of shales, coals and sandstones of the Inverness Fm (Cumberland Gp) into moderately dipping attitudes adjacent to the diapir. (From Alsop et al., 2000)



Figure 6. Photograph taken on trip looking north to the white gypsum mylonites of the Finlay Point Diapir (on the right of the photograph) in tectonic contact with conglomerates, sandstones and coal seams of the Inverness Fm. The NW-dipping overburden displays a progressive reduction in dip away from the diapiric contact over a distance of 250 m. (Interpretation from Alsop et al., 2000)

Displacement across the major extensional fault bounding the inner drag zone can not be precisely determined, although the lithological variation across this fault suggests significant movement (>100 m). Outwards of this inner fault-bounded drag zone, bedding in the Inverness Fm typically dips at c. 25°, reducing to 20° at greater distances from the diapir. Two sets of minor conjugate extensional fractures cut bedding throughout the drag zone resulting in significant bedding-parallel extension. These fracture sets display a constant parallel or radial arrangement to the diapiric margin despite variations in fault density and strain through the drag zone. The 300 m wide drag profile is transected by major NNE trending steeply west-dipping late-extensional faults with down-dip slickenlines. Individual coal seams can be correlated across fault planes indicating displacements of <5 m (Fig. 5c). Bedding-parallel slip occurs within the coal-shale-sandstone succession due to the heterogeneous nature of the sequence, with coal seams recording the highest fracture density measurements and concentrating deformation despite being further removed from the diapiric margins than the conglomerate. Early extensional faults may be rotated adjacent to the diapir, resulting in apparent contractional offsets (Fig. 5c).

Coal Mine Point Diapir

The Coal Mine Point Diapir and its associated drag zone are exposed in a 1 km section on Mabou Mines beach (Fig. 8), where the shales, coal seams and sandstones of the Inverness Fm form the western margin to the diapir (46°07'29"N 61°27'55"W) (Figs 1b, 7a, b). Seismic data indicate that the exposed drag zone is located on the steeply dipping flank of the diapir, with the Hollow Fault forming the SE-margin (Fig. 7b). The 250 m wide drag zone is developed in shale-dominated Inverness Fm. Braided fluvial sandstones (Eagle Sandstone) up to 93 m thick are surrounded by the incompetent shales and coals thus providing a marked lithological heterogeneity. The extremely high competence contrasts developed between predominant shales and sandstones hinders the development of a rigid sandstone framework to the drag profile. The deformation within the drag zone is fundamentally different from that observed at Finlay Point. The margin of the diapir is marked by a tightly folded mylonitic gypsum foliation containing a gently plunging fabric intersection lineation which is parallel to the diapiric margin (Fig. 7c). Internal fracturing within the gypsum results in gently dipping gypsum veins with vertical gypsum fibre infill. In the shales adjacent to the contact, scaly clay fabrics are developed sub-parallel to bedding whilst sandstones and thin limestones contain sub-vertical calcite veins. A marked 250 m wide drag zone is developed with SW dipping sandstones and shales undergoing a progressive (57°) rotation to steeply dipping attitudes adjacent to the diapir, whilst maintaining NW–SE strikes. The 43 m thick, relatively homogeneous Stack Sandstone has been reduced to a breccia adjacent to the diapir, with pervasive NNW–SSE-trending, steeply east-dipping minor extensional faults and sub-vertical granulation seams displaying brittle fracturing but no dynamic grain size reduction. Inward-dipping extensional faults produce a thinning of the competent sandstone immediately adjacent to the shoulder of the diapir (Fig. 7c). Intervening packages of shale (<5 m thick) and coal seams show abundant evidence of bedding-parallel shear, with the development of bedding-parallel fractures and slickensides, polished surfaces and anastomosing shear zones with scaly clay textures (Fig. 7c). A major N–S

trending bedding sub-parallel fault dips steeply towards the west within the shales and coals. (Fig. 7c). The shales and coal seams are therefore considered to be relatively incompetent and to absorb a large proportion of the overall deformation by a process of bedding-parallel shear and flexural flow which is equivalent to faulting in more competent units. The pervasively fractured sandstones have deformed at higher strain rates close to the diapiric margin, with strain being localized in bedding-parallel shears in the overlying shales which thus protect and cushion the Eagle Sandstone from greater deformation.

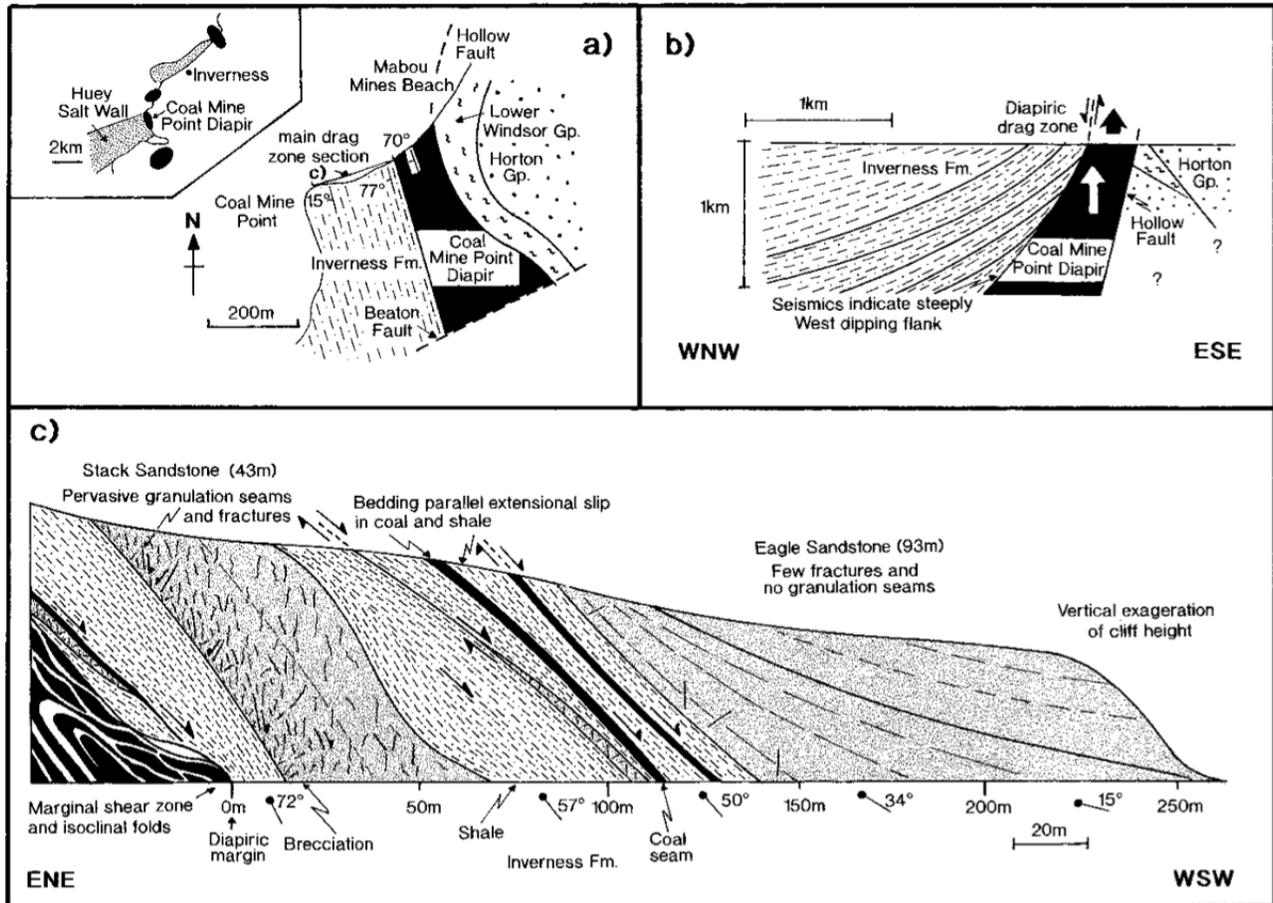


Figure 7. (a) Simplified geological map of the Coal Mine Point Diapir and adjacent areas (modified from Giles et al. 1997a, b). The map highlights the position of the main coastal drag zone section shown in (c), whilst the inset map shows the general location of the diapir (refer to Fig. 1b). (b) Schematic WNW–ESE section illustrating the overall geometry of the Coal Mine Point Diapir based on outcrop studies and offshore seismic data. The relative location of the detailed diapiric drag zone section (c) is noted. (c) ENE–WSW section through the Coal Mine Point Diapir drag zone showing rotation of shales, coals and sandstones of the Inverness Fm (Cumberland Gp) into steeply-dipping attitudes adjacent to the diapir (From Alsop et al, 2000).

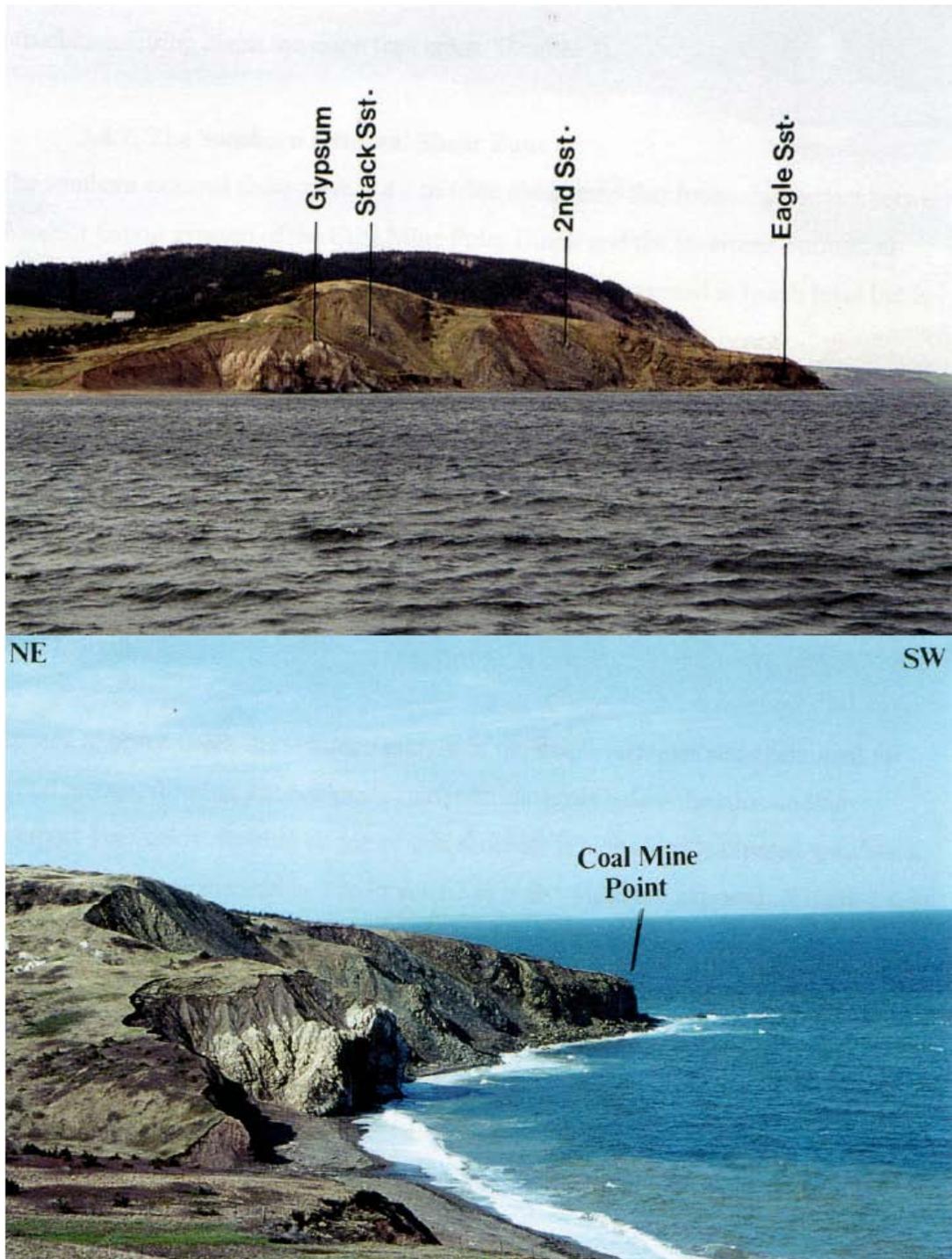


Figure 8. Photographs of Coal Mine Point Locality. (Brown 1998)

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Ocean Patterns around Nova Scotia

Contributed by Chen Chen

Nova Scotia's ocean circulation and tidal features will be discussed in this section. Figure 1 shows the circulation patterns of the three bodies of water that surround Nova Scotia: the Northern Atlantic to the southeast, the Gulf of St. Lawrence to the north, and the Bay of Fundy and Gulf of Maine to the southwest.

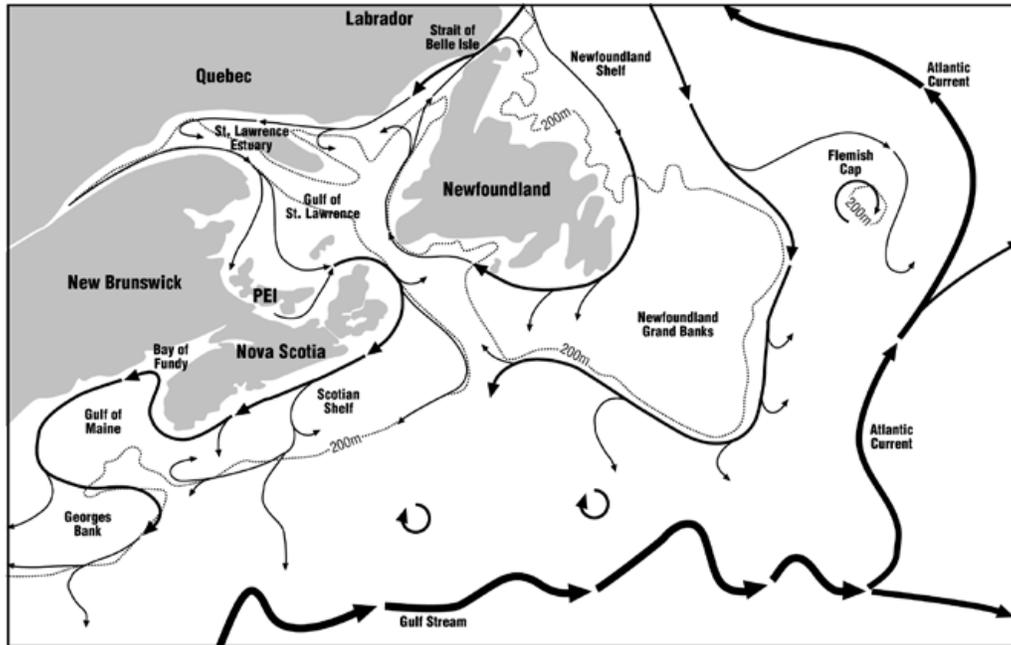


Figure 1: Oceanic circulation pattern around Nova Scotia, from Davis and Browne (1996).

Several hundred kilometers south of Nova Scotia, the Gulf Stream carries a prodigious amount of heat and water northward (approximately 50 million m^3/s), which influences the large-scale climate over Atlantic Canada. The Gulf Stream is the strong (up to 2 m/s) western boundary current that forms the clockwise subtropical gyre in the Atlantic Ocean, which is driven by the basin-scale wind stress curl. The current divides near eastern edge of the Newfoundland Grand Banks. While a part of it continues NE as the North Atlantic Current, the remainder turns southeast to complete the subtropical gyre and join back with the North Equatorial Current.

The Labrador Current flows SE along the coasts of Labrador and Newfoundland (approximately 5 million m^3/s). It branches off Newfoundland, the main portion flowing along the slope of Grand Banks and the minor part flowing inshore over the shelf. This inshore part later divides to provide part of flow through the Strait of Belle Isle while the remainder flows along the eastern and southern coasts of Newfoundland and joins the Gulf of St. Lawrence circulation through Cabot Strait.

The Gulf of St. Lawrence is a highly stratified semi-enclosed marginal sea. Circulation in the Gulf is influenced by freshwater runoff originating from the St. Lawrence River and North Shore rivers, local winds, and tides. The surface circulation in the Gulf of St. Lawrence is essentially counterclockwise. In the vicinity of the Cabot Strait (the southern entrance of the gulf between Cape Breton and Newfoundland), the fresh-water flow (less dense) exits the Gulf on the surface layer, which is then compensated by a salt-water flow (more dense) that enters from the continental shelf through a lower layer.

After the fresh water rushes out from the Gulf, the Coriolis effect from Earth's rotation turns the flow to the right. Instead of rushing toward the deep ocean, the freshwater export out Cabot Strait becomes the Nova Scotia Current, flowing southwesterly along the Nova Scotian Atlantic coast. Its annual mean transport is approximately

0.4 million m^3/s , with the current speed of 0.05– 0.10 m/s. The coastal current is in quasi-geostrophic balance, with the pressure gradient to the left balanced by the Coriolis force to the right. Thus, we observe flow that mainly follows the topographic contour, or isobar.

The Nova Scotia Current then follows the coast, entering the Bay of Fundy and Gulf of Maine. There is a counterclockwise surface circulation, which brings Scotian Shelf water westward into the Gulf and contributes to the eastward jet (0.3 m/s) along the inner edge of Georges Bank. Since the tidal current is very strong in the Gulf, “circulation” here is actually the residual current of the tidal back and forth motion.

The ebb and flow of tides are the consequence of a combination of the gravitational force (from the moon-sun-earth system) and the centrifugal force (resulting from the revolution of the moon about the earth as well as the revolution of earth about the sun). The motion of the moon and sun are not synchronized; thus, a tidal range can become larger during a spring tide, when the sun, earth, and moon are aligned resulting in a full (or new) moon. The tidal range is the smallest during a neap tide, when the sun, earth, and moon are at a right angle resulting in a half moon. The observed tidal response can be decomposed into different components. Usually, the largest component turns out to be the principal lunar semi-diurnal M2: high water level occurs twice per day, with 12.4 hours interval. There are also solar semi-diurnal components, lunar and solar diurnal components. Superimposed upon the astronomical tides are meteorological tides resulting from anomalous winds and atmospheric pressures. Generally, the tidal periods and the variation between spring and neap tide can be explained by the equilibrium tide theory with the assumption of an aqua planet model. In reality, tidal periods are constrained by ocean basins, which produce a more standing wave pattern than a progressive wave pattern. The rotation of the earth also distorts the horizontal component of a tidal flow. These effects result in an amphidromic tidal system (Figure 3a). This tidal system requires a node with no amplitude and that tides propagate around this node. The cotidal line indicates the same phase while the corange line indicates the same amplitude. Tides belong in a class of long gravity waves; thus, the wave speed is proportional to the square root of the depth. For example, in the open ocean (3 km depth), the wave speed is approximately 170m/s. For the less deep basins surrounding Nova Scotia, the wave speed is far less: 45 m/s in the Gulf of St. Lawrence (0.2 km) and 30 m/s in the Bay of Fundy (0.1 km).



Figure 2: Low to high tide at one harbor near the head of Bay of Fundy.

The Bay of Fundy is known for having the highest tidal range in the world, with Burntcoat Head, Nova Scotia officially regarded as having the highest tides in the world by The Guinness Book of World Records (1975) (the greatest mean spring range of 14.5 meters and an extreme range of 16.3 meters). Besides Bay of Fundy, other marginal seas like Ungava Bay in northern Quebec, King Sound in Western Australia, Gulf of Khambhat in India, and the Severn Estuary in the UK, also exhibit high vertical tidal ranges (>10m). Three photos in Figure 2 (captured during the 2012 Nova Scotia field trip) show the low to high tide near the head of Bay of Fundy. Tons of water retreat from and rush into the mudflat every 12.4 hrs, dominated by Principal lunar semidiurnal constituent. Figure 3d shows the cotidal line and corange line in the bay, and 3c shows a simplified model for this long (150km) and shallow (90m) marginal bay. The width of the bay (50 km) is much less than the Rossby deformation radius (400 km at this latitude); therefore, the rotation of the earth does not have large effect within the Bay of Fundy.

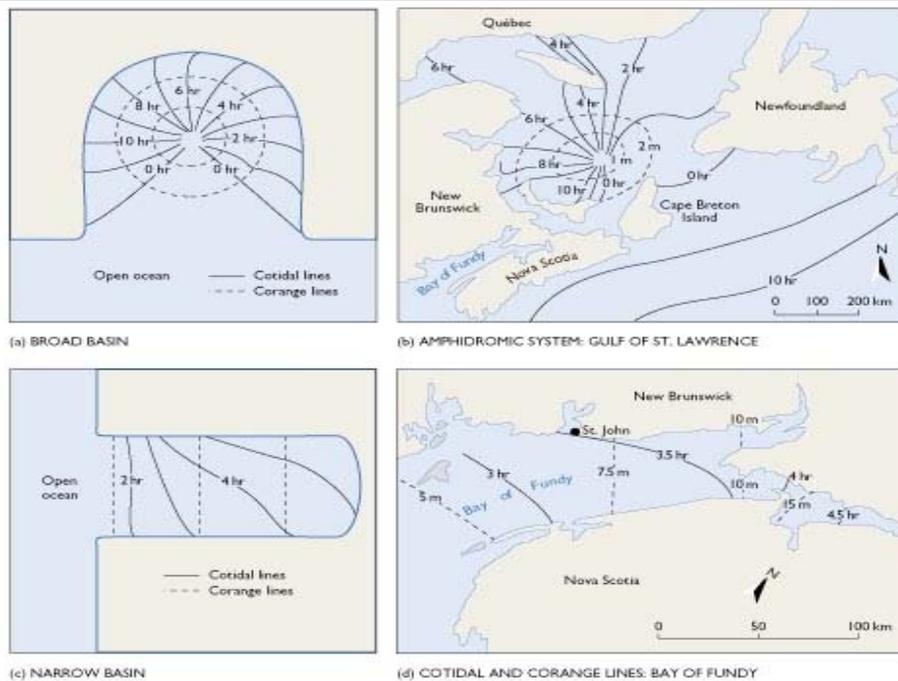


Figure 3: M2 tides in Gulf of St. Lawrence and Bay of Fundy, from Pinet (2012).

On the contrary, the tides in Gulf of St. Lawrence are a combination of semi-diurnal and diurnal constituents with ranges from 1.1 to 2.9 m. One may wonder why these two neighboring bodies of water have such a different tidal system. The width of this Gulf (400 km) and depth determines the magnitude of the tidal range. Even though the Gulf is large enough to observe an amphidromic system (Figure 3b), this is inconsequential to the overall tidal range.

It is well known that the average open ocean tidal range is around 1 m and marginal seas nearly 2 m. One can ask why so few places worldwide can generate such a large tidal range (> 10 m) as observed at the Bay of Fundy. The mechanism for these tidal anomalies is the tidal resonance (Garrett, 1972; Godin, 1993). In the case of the Nova Scotia, the 12.5 h period of the lunar semi-diurnal tidal driving force coincides fairly close with the intrinsic resonant period of the Gulf of Maine/Bay of Fundy system, which is determined by the depth and length of the system. This explains why some regions with a similar size can produce a larger tidal range in comparison to others. Usually, shallow (<100 m) and long (200 km) semi-enclosed basins or seas have an intrinsic oscillation period near 12 hr, which can be triggered by the forced tidal motion. These unique systems experience resonance with the M2 tide of the adjacent open ocean and produce much higher tides in the bay in comparison to the deep ocean.

Some have suggested building a tidal power station in the Minas Basin; however, computer model studies (Greenberg, 1983) of the Bay of Fundy tides suggest that a tidal barrage (i.e., a dam for tidal power) placed across the Minas Basin would result in the shortening of the Bay. This would effectively bring the natural period of the Bay of Fundy/Gulf of Maine system closer to a forcing tidal period. The tidal amplitude would be reduced in the Minas Basin, but as a result the tidal amplitude from Saint John to Boston would increase by at least 15 cm. This result of this modeling study serves as a reminder that we need to consider and minimize the environmental effect (both local and remote) before we build any facility to take advantage of the large tidal range in Bay of Fundy.

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Tidal Bore and Macrotidal Sedimentary Structures in the Bay of Fundy

Contributed by Amelia Paukert

The Bay of Fundy has the highest tidal range in the world, with an average maximum tidal range of 13.25m (Archer & Hubbard, 2003). The tidal range increases as you go inland from the bay, as the water channel narrows and shallows (see Figure 1). As the incoming high tide drives large volumes of water inland, there is less available volume in the channel to accommodate the water, forcing the water level higher.

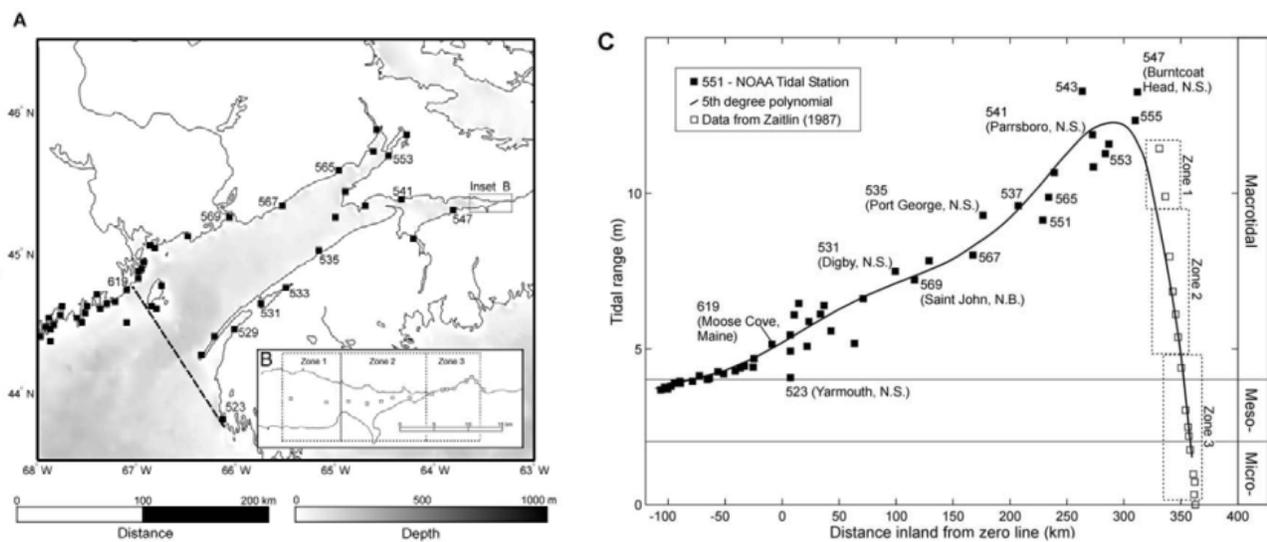


Figure 1: a) Bay of Fundy map with locations of tidal range measurements marked, c) Tidal ranges for locations in the Bay of Fundy from the entrance of the bay up to the point of Cobequid Bay in the Minas Basin.

Tidal Bore

Preceding high tide, the front of water rushing into restricted areas such as rivers and narrow parts of the bay forms a tidal bore. As described in Archer and Hubbard (2003), tidal bores can be broken down in the following way:

“The tidal bore is composed of three parts: the bore itself, the turbid zone, and the whelp (see Figure 2). The bore is the main front of the arriving water and looks like a breaking wave in shallow water. In areas of the Bay of Fundy, the bore can reach heights of up to 2m. The turbid zone is a narrow finger along the edge of the waterway that extends in front of the main bore and scours the underlying sediment of the lower intertidal mudflats. The turbid zone has high viscosity due to its high suspended load. The whelp is characterized by closely spaced waves that indicate the presence of the bore in deeper water.”

The tidal bore is generally fresh water and precedes the arrival of the flood tide (the rapid rise in water level just before high tide) by about 25 minutes (Archer, 2004). Higher salinity water arrives about 10 minutes after the flood tide.

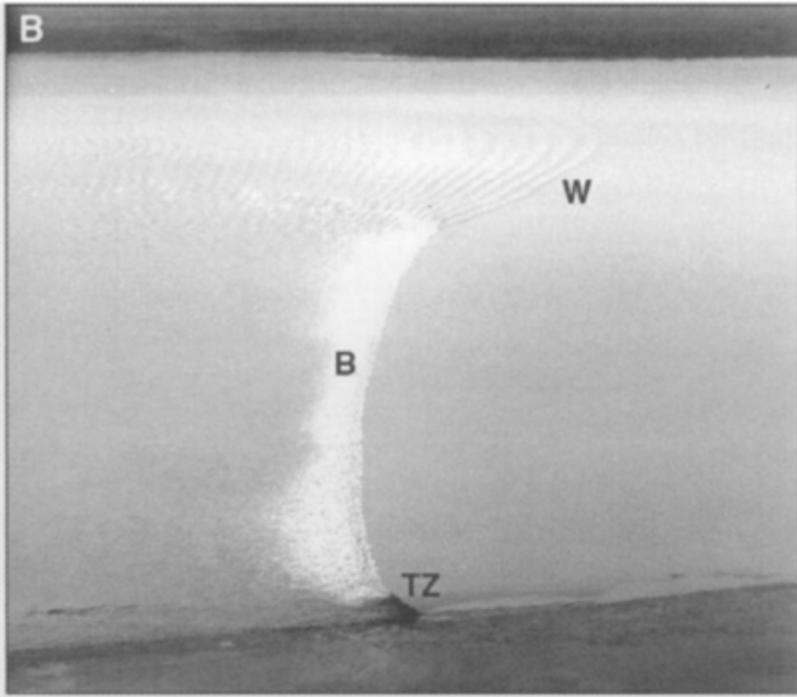


Figure 2: Diagram of parts of the tidal bore. TZ indicates the turbid zone, B the tidal bore, and W the whelp. From Archer (2004).

Macrotidal Sedimentary Structures

The Bay of Fundy is a macrotidal region, and has well-preserved sedimentary structures. Macrotidal areas are characterized by high rates of sedimentation, as well as low biological activity and consequently low rates of bioturbation.

The fast rate of tidal water flow (1-2.5 m/s) leads to high turbidity³. The low intertidal zone is dominated by erosion, with pitting and scouring. The upper intertidal is dominated by deposition, with low energy flooding forming laterally extensive mudflats. There is an overall landward migration of sediment.

For a more detailed look at the depositional zones, the Cobequid Bay in the Minas Basin can be broken into three main parts (see Figure 3): Zone 1 (Cobequid Bay), Zone 2 (intermediate between the bay and the river), and Zone 3 (the Salmon River).

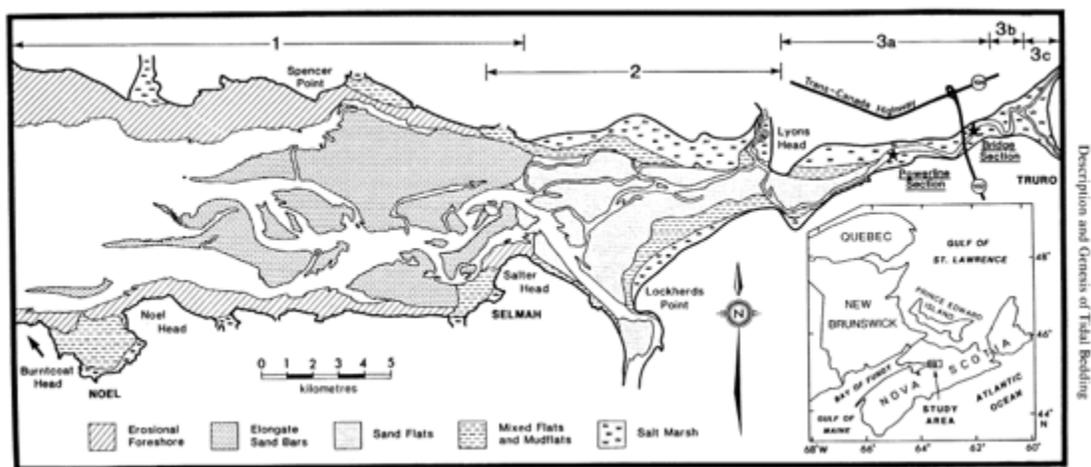


Figure 3: Zones of deposition in the upper Minas Basin. From Dalrymple and Makino (1989).

Zone 1, e.g., Cobequid Bay, has elongate tidal sand bars with dunes and scour pits (Figure 4a). Zone 2 has upper flow regime sand flats with current ripples overlying parallel lamination (Figure 4b). Zone 3 is a tidally influenced confined channel, with steep erosional scarps from channel migration at the edge of salt marshes (Figure 4c).



Figure 4, from Dalrymple and Makino (1989): left, dunes and scour pits; middle, current ripples; right, erosional scarp

Dalrymple and Makino (1989) describe macrotidal sedimentation in the following manner:

“The mudflats have exceedingly high rates of sedimentation, with 1-13mm of deposition each tidal cycle. The formations from this fast sedimentation are called cyclic rhythmite, because they alternate layers of sand with layers of mud (see Figure 5). The sand is deposited during periods of transition between tides, while the mud is deposited during the static period of high or low tide. The thickness of the sand layers depends on the strength of the tide, while the mud layer thickness is determined by the height of the river, so it is generally constant. Rhythmite has 4 main structures: flood tide ripples, ebb tide ripples, upper flow regime parallel laminations, and suspension deposits of thin muddy sand. The speed of water flow determines which structures are formed, with suspension deposits forming in slow flow, ripples forming in intermediate flow, and parallel laminations forming in areas of high flow. In a particular location there is a muddying upward trend as the accommodation fills in with new material and the channel shifts to another location.”

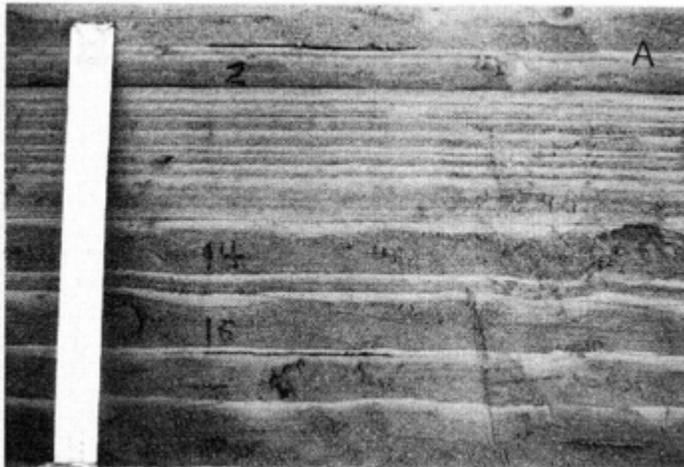


Figure 5: Tidal cyclic rhythmite, from Dalrymple and Makino (1989).

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V. Detailed Itinerary and Descriptions of Sites Visited

Day 1 (August 23rd, 2012): Hartford Basin (Eastern Daylight Time)

by Jesse Farmer

- 9:00** Leave Lamont-Doherty Earth Observatory
- 11:10** Arrive Outcrop 1: New Haven Formation Roadcut at I-84, I-691 junction. Spend 30 minutes before being told to leave by State Police.
- 12:00** Arrive Outcrop 2: Talcott Basalt (Meriden Target)
- 13:00** Depart Outcrop 2
- 13:20** Arrive at Dinosaur State Park. Eat lunch and tour the Park.
- 14:50** Arrive at Outcrop 3: East Berlin Formation (Berlin Turnpike)
- 15:50** Arrive at Outcrop 4: Portland Formation (Manchester, CT Home Depot)
- 16:30** Cars split up and depart for Lemoine State Park, Lemoine, ME. Pick up dinner along the way.
- 23:00** Arrive **Lemoine State Park**, Lemoine ME. Set up camp in the rain.

Site Descriptions

Stop 1: New Haven Formation, 41.559 N -72.907 E

The New Haven Formation is a ~2 km thick sequence of intensely bioturbated mudstones and soil carbonate layers consistent with a large braided river system continually filling available accommodation space. Paleo-latitude studies of the New Haven Fm places it at approximately 15°N, within the tropics. Uranium-lead dating of the conspicuous soil carbonate layer at this outcrop (Figure 1a) dates to 211.9 ± 2 Ma (late Triassic), and paleosol CO₂ reconstructions places atmospheric CO₂ at 1000-2000ppm during this interval. Rhizomorphs and rhizoliths (traces of root activity) are present to abundant in the mudstones.

Stop 2: Talcott Basalt 41.553 N -72.816 E

Exposed directly in the outcrop behind the Meriden Target are lacustrine sequences of the upper New Haven Fm overlain by ~20 m of pillow basalts of the Talcott Basalt (Figure 1,c). The pillowed textures grade into more massive and columnar basalts toward the far end of the outcrop. The Talcott Basalt is associated with the Central Atlantic Magmatic Province and may have accumulated at rates of up to 1 m per year.



Figure 1. (a) soil carbonate layer of the New Haven Fm; (b) uppermost New Haven Fm (units at bottom) overlain by Talcott Basalt pillows; (c) Talcott Basalt outcrop; note large brown pillow directly above poster.

Step 3: Dinosaur State Park

Abundant fossil three-toed *Eubrontes* tracks are preserved in mudstone layers of the upper East Berlin Formation at Rocky Hill, CT (Figure 2a). Originally excavated as the site of a state building, the site was preserved as a state park in 1966.

Stop 4: East Berlin Formation 41.622 N -72.739 E

Over 120m of Early Jurassic lacustrine reddish sandstones, yellowish siltstones and gray-black laminated claystones are exposed along the roadcuts at the intersection of Route 9 and Route 5-15. These units are attributed to the upper East Berlin Formation and are overlain by the Hampden Basalt, which is visible up-section along the CT-9 roadcut. The gray-black laminated claystones contain generally abundant fish fossils and suggest a depositional environment in deep lake with anoxic sediments (Figure 2b). Three units of gray-black claystones are apparent along this outcrop, which have been interpreted to represent three Van Houten cycles with approximately 21,000 years separating each lake highstand (precessional insolation forcing).

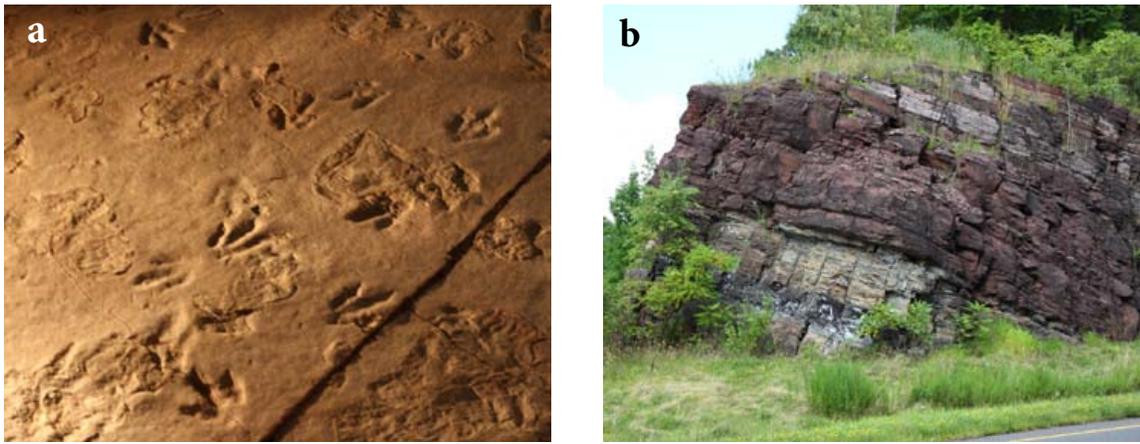


Figure 2. East Berlin Formation. (a) track impressions at Dinosaur State Park. (b) a Van Houten cycle at the CT-9 road cut, showing gray-black shales (deep lake) grading into yellowish siltstones and red sandstones (shallow lake/fluval).

Stop 5: Portland Formation 41.806 N -72.553 E

Jurassic-age fluvial deposits of coarse, feldspathic arenite are exposed in roadcut along Buckland Hills Dr. This is the site of the old Portland Brownstone Quarries, from which sandstone blocks used to build many brownstone buildings in New York City, the Klein Geology building at Yale University, and academic buildings of New York University were sourced. Eight nearly complete dinosaur skeletons were discovered in the Portland Fm in the 19th century. Interestingly, there is almost no soil carbonate in the Portland Fm, in contrast to the New Haven Fm. The paleo-latitude of the Portland Fm is subtropical (~40°N), where arid conditions leading to abundant soil carbonates would be expected. With CO₂ drawdown from weathering of CAMP basalts, however, the climate of the Jurassic during the deposition of the Portland Fm would be much cooler than the Triassic.

Day 2 (August 24th, 2012): Maine Ice Age Trail

by Alison Hartman and Mike Wolovick

- 7:30 Wake up, breakfast
- 9:00 Depart Lemoine State Park Campground; Lemoine, ME
- 10:30 Long Cove Picnic Area (Ice Age Trail #10)
- 11:30 Ridge Road (Ice Age Trail #17-19)
- 13:00 Lunch at Columbia Falls, ME
- 14:20 Quarry on Rt. 191 (Jacksonville Esker; Ice Age Trail #32)
- 14:30 Continue travel into Canada

Site Descriptions

Introduction

The Maine Ice Age Trail (<http://iceagetrail.umaine.edu/index.php>) was developed by the University of Maine and is comprised of stops featuring excellent examples of features made during the presence and retreat of the Laurentide Ice Sheet. This ice sheet covered the entirety of the state and a large portion of northern North America during the last glacial maximum (25,000 kyr B.P.). A main feature seen along the route is evidence for local sea level change associated with the rebound of the land in response to the absence of the ice sheet. This process, known as post-glacial rebound, means that most of the geomorphology of the region was actually formed in a marine environment and only now are exposed above sea level. This process continues to this day and there are several historical stops along the trail that show the results of local sea level change (such as the remains of piers that are now some distance from the shore). The migration of salt marshes in the area also demonstrates these changes. The main features of the trail are eskers, moraines, deltas and glacial erratics. These features are made up of unsorted glacial till and as a result are especially ideal for blueberry farming. All stops along the trail are numbered and labeled on site. A copy of the map can be ordered through the Maine Ice Age Trail website along with a complete description of each stop.

Stop 1: Long Cove Picnic Area (#10)

- Glacial striations on granitic bedrock show the direction of the ice flow (oriented $\sim 150^\circ$)
- This site experiences large tidal changes and during low tide there is an expansive mudflat that consists of marsh grasses, seaweed, crabs and multiple other organisms.

Stop 2: Ridge Road (#17-19)

- Road runs along crest of terminal moraine, driving east.
- Along an unmarked dirt road (left off Ridge Road), there are several cuts along the moraine (Figure 1a)
 - Unsorted till deposits exposed near an abandoned trailer
 - Exposure ~ 5 m high
 - Numerous ~ 1 m erratic boulders
 - Boulders: granite/diorite
 - There are a series of smaller moraines subparallel to the main terminal moraine.
- Back on Ridge Road: boulder field on either side of road. Numerous erratics 1-3 m scale (Figure 1b).

Stop 3: Quarry on Rt. 191 (Jacksonville Esker, #32)

- Rt 191 runs along crest of an ~ 30 m tall esker for ~ 3 miles
- Visible unsorted till (mostly sand and gravel with some erratics) at quarry location



Figure 1. (a) Ridge Road moraine cut; (b) erratics in blueberry fields.

- 8:00 Wake up
- 8:30 Breakfast
- 9:30 Drive from Seaside Tent & Trailer Park to The Caves Restaurant on Big Salmon River Rd (2 km)
- 9:35 Explore the caves of St. Martins (only accessible at low tide)
- 10:30 Drive from St. Martin's Beach to the Fundy Trail Parkway (11 km)
- 10:40 Enter Fundy Trail Parkway, and go to stop B, Melvin Beach.
- 10:45 Hike down to and explore Melvin Beach
- 13:20 Leave Fundy Trail Parkway and return to The Caves Restaurant
- 13:30 Lunch
- 14:15 Leave the Caves and start heading towards Five Islands
- 15:00 Stop at km 184 marked on Rt. 1; explore roadcut of lacustrine shale
- 15:45 Continue on towards Five Islands
- 17:30 Arrive at **Five Islands Provincial Park**

Site Descriptions

St. Martin's Beach

St. Martin's Beach is impressive not only for the sea caves accessible here at low tide, but for a nice outcrop of the Carnian unconformity, separating tectonostratigraphic units I (fluvial Honeycomb Pt. Formation sandstone, right) and II (fluvial Quaco conglomerate, left) (Figure 1a,b). The former dips 2-3° more steeply than the latter, "consistent with tilting towards the border fault complex prior to the deposition of the Quaco" (Olsen and Touhami, 2008). The clast-supported texture and presence of organically preserved plant materials suggests a significantly more humid depositional environment for the Quaco. Quartzite clasts are spotted with rough pale craters, which when examined in situ apparently form at contact points with other clasts. While recent interpretations have suggested the passage of a shock wave from extraterrestrial impact as the source of these craters, they are

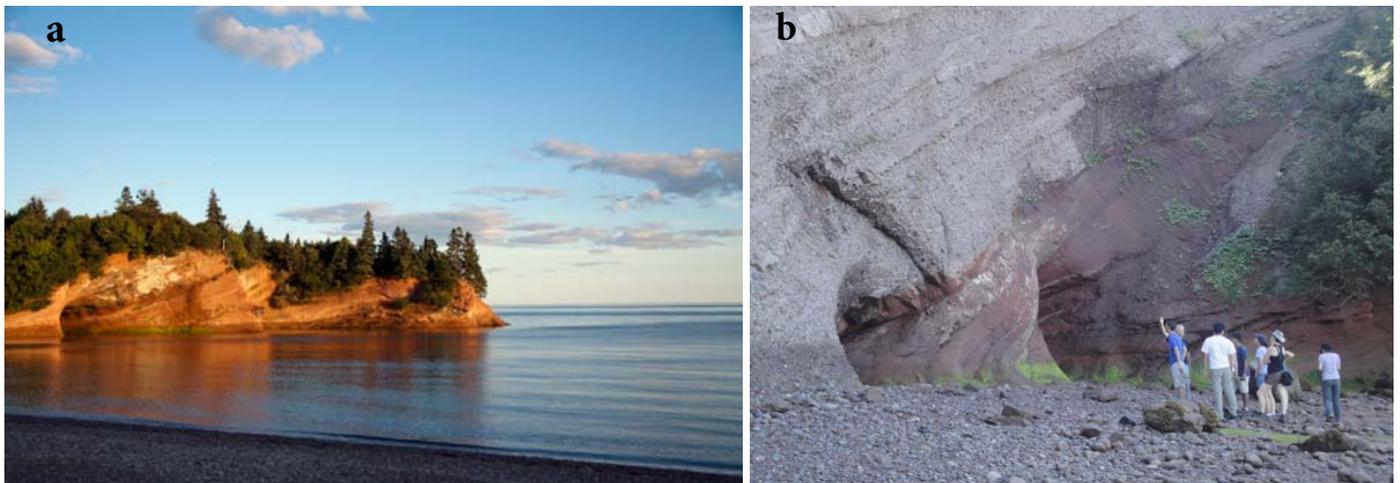


Figure 1. St. Martin's Beach. (a) At high tide; (b) Carnian unconformity separating TS-I (reddish Honeycomb Pt. Fm., bottom) from TS-II (grayish Quaco Conglomerate, top).

more commonly interpreted as being due to pressure solution (Olsen and Touhami, 2008). The caves themselves, apart from demonstrating the results of tidal erosion, contain nicely preserved sedimentary features, particularly desiccation cracks and ripple marks, and sporadically quadrupedal footprints (Olsen and Touhami, 2008).

Melvin Beach

Melvin Beach (Figure 2) is flanked to the east and west by cliffs of the Echo Cove Member of the Wolfville Formation. This member overlies the Quaco conglomerate and “consists mostly of gray, tan, and red sandstone and gravelly sandstone with less abundant red and gray mudstone,” indicative of deposition by a system of small streams as opposed to the large rivers associated with the Quaco (Olsen and Touhami, 2008). Fownes Head on the west side of the beach contains well preserved plant fragments and wood, and has produced pollen and spores indicating a Carnian age. The red clastic rocks on the east side of the beach are another subdivision of the Echo Cove Member, and have been interpreted as overlying the Fownes Head member, although this relationship cannot be demonstrated here (Olsen and Touhami, 2008).



Figure 2. Echo Cove Member of the Wolfville Formation, showing cross and interbedding with large changes in grain size and sorting.

Albert Formation

The Albert formation is a Mississippian lacustrine sequence within the Moncton basin, a transpressional basin formed during the docking of African and North American plates during the Carboniferous. Abundant organic rich rocks contain assemblages of fish and plant fossils, and have sourced hydrocarbon reservoirs in sandstones of the same formation, resulting in two producing commercial fields in the area: Stoney Creek gas and oil field south of Moncton produced 28.7 bcf of gas and 800,000 barrels of oil between 1909 and closure in 1991, and the McCully gas field, discovered in 2000 with proven and probable reserves estimated at 119 bcf (Olsen and Touhami, 2008).

Reference

Olsen, P., and M. Et-Touhami. 2008. “Field Trip 1: Tropical to Subtropical Syntectonic Sedimentation in the Permian to Jurassic Fundy Rift Basin, Atlantic Canada, in Relation to the Moroccan Conjugate Margin”. Central Atlantic Conjugate Margins Conference, Halifax, Nova Scotia, August 2008. ISBN:0-9810595-3.

Day 4 (August 26th, 2012): Joggins Fossil Cliffs, Cape Chignecto & Clarke Head by Anna Foster

- 7:30 Wake up
- 8:40 Depart Five Islands Campsite
- 9:45 Arrive at Joggins Fossil Cliffs
- 10:00 Museum Tour
- 11:00 Beach Tour
- 12:30 Lunch
- 13:15 Depart Joggins
- 14:15 Arrive Cape Chignecto Park, walk Beach Trail
- 16:00 Leave Cape Chignecto Park
- 17:10 Arrive Glooscap Campground to see Clarke Head
- 18:30 Depart Glooscap Campground
- 19:00 Return to **Five Islands Provincial Park**

Joggins Fossil Cliffs Museum/Field Observations

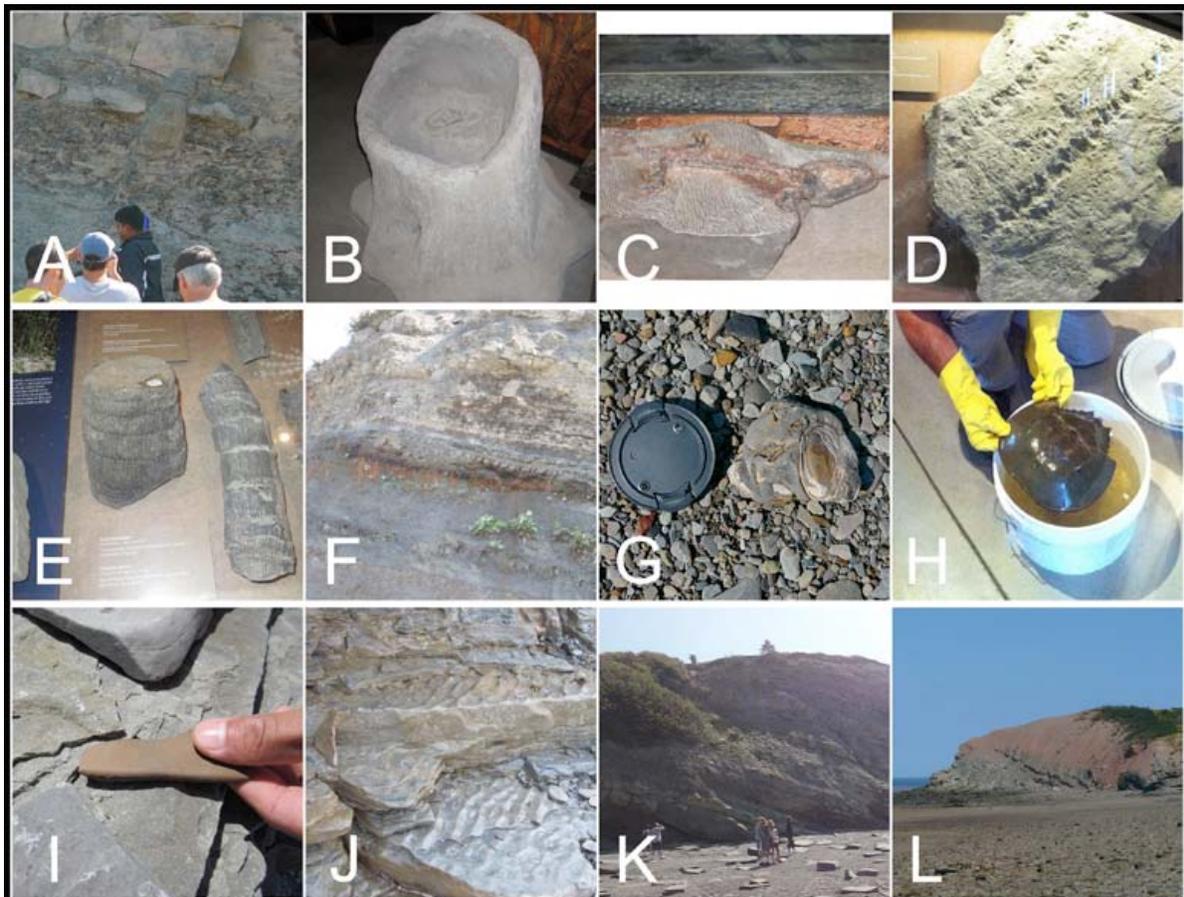


Figure 1. Photographs of museum and field observations from Joggins. Explanations on following page.

The Joggins section itself contains sandstone and mudstone red-beds, dominating in the lower layers, alternating with limestone and coal, which dominate the upper layers and are associated with the base level rhythms discussed in the Geologic Overview. Sandstone channels can be observed in the red-beds, resulting in unconformities with very short erasure, from months to a few years. The limestone/coal deposits contain ‘fish poop fossil’ and small bones, indicating a relatively turbulent deposition. ‘Clam coal’ is also observed. The present-day tilt of the strata is due mostly to salt movement, as well as later tectonics. Fossil preservation of bones at Joggins is usually in the form of mineralization with silica, because the plant-rich environment of the rainforest would have resulted in low pH conditions, encouraging silica precipitation. This has the added benefit that approximately 80% of the time, both the cast and mold of the fossil are recoverable (Joggins Fossil Cliffs Tour Guide, personal comm., 2012).

Captions for Figure 1:

- A. Fossilized piece of a *Lycopod*, a spore-producing plant-like tree that grew to 30—40 m tall. An important factor in determining the in-situ nature of fossils at this site was the fact that this and other fossilized trees are upright, and the surrounding sedimentary layers are undisturbed.
- B. Museum representation of *Hylonomus lyelli*, the first amniote/first reptile that could lay eggs on land. These fossil reptiles were discovered inside the fossilized *Lycopods*, adding further evidence to the characterization of Joggins as a parautochthonous assemblage.
- C. *Dendroperpeton acadianum* fossil, a primitive amphibian, estimated to be 315 ma. This specimen was discovered in 1986 by two students from McGill University (Godfrey et al., 1987), and is considered the best fossil at the Joggins museum.
- D. *Arthropleura* trackway- the track is approximately 5 m long, and was discovered in 1963. It was made by *Arthropleura*, a millipede that was about 6ft long and 1 ft wide, making it the largest known land invertebrate.
- E. Fossil *Calamites*, a segmented, grassy, spore-bearing plant that grew to be 10 m high. It is closely related to modern-day horsetails.
- F. Coal seam, smaller than used for commercial mining, but likely of the type used by native and early settlers as fuel.
- G. Example of Joggins coal, also called clam coal, because it contains lots of peat, fish scales, clams, and other recognizable organic material.
- H. Modern-day preserved horseshoe crab, approximately 1 ft across, and very similar to horseshoe crabs from 350 ma, which were the size of a half dollar.
- I. An iron-ore nodule, formed in the tropical rainforest environment of the Pennsylvanian when iron evaporated, rained, and ran down grooves in the bark of the *Lycopod* trees, creating very smooth nodules at the base of the grooves in the bark.
- J. Fossil ripples, mostly symmetric but with a paleocurrent direction visible. Load structures were also observed in the layers (not pictured).
- K. Section of tilted strata, representative of the entire section and, combined with tidal erosion, the reason the Joggins fossils are so accessible.
- L. Glacial till (red strata) overlying the Joggins formation, deposited approximately 10 ka.

Cape Chignecto Park

Cape Chignecto splits the Fundy Basin into the Chignecto Bay to the north, and the Minas Basin to the south. It is encompassed by Cape Chignecto Provincial Park, the largest provincial park in Nova Scotia and an important wilderness reserve. The cape marks the western edge of the Cobequid Hills, and the Cobequid fault segment is found offshore (Figure 2). To access the site, park and sign in (and pay) at the Cape Chignecto Provincial Park Visitor’s Center. Take the beach trail; this is a very short and easy walk to the water.

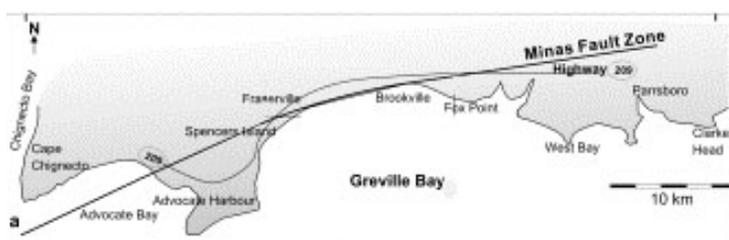


Figure 2. Map of Cape Chignecto (west) and Clarke Head (east), with the Minas Fault Zone (also called the Cobequid fault segment) shown in black (Waldron et al., 2005).

On the beach, immediately to the east (left) are large outcrops of red sandstone. These are of the lower part of the Blomidon Formation. There have been reports that some of these are Eolian in origin, although we did not see evidence of this. Bioturbation, channels, rip-up clasts, fractures, and burrows in the mudstone were observed (Figure 3). Additionally, there is an apparent 'vertical debris flow', in which unsorted grains, including larger grain sizes than are present in the surrounding layers, cut vertically through many horizontal layers (Figure 4).

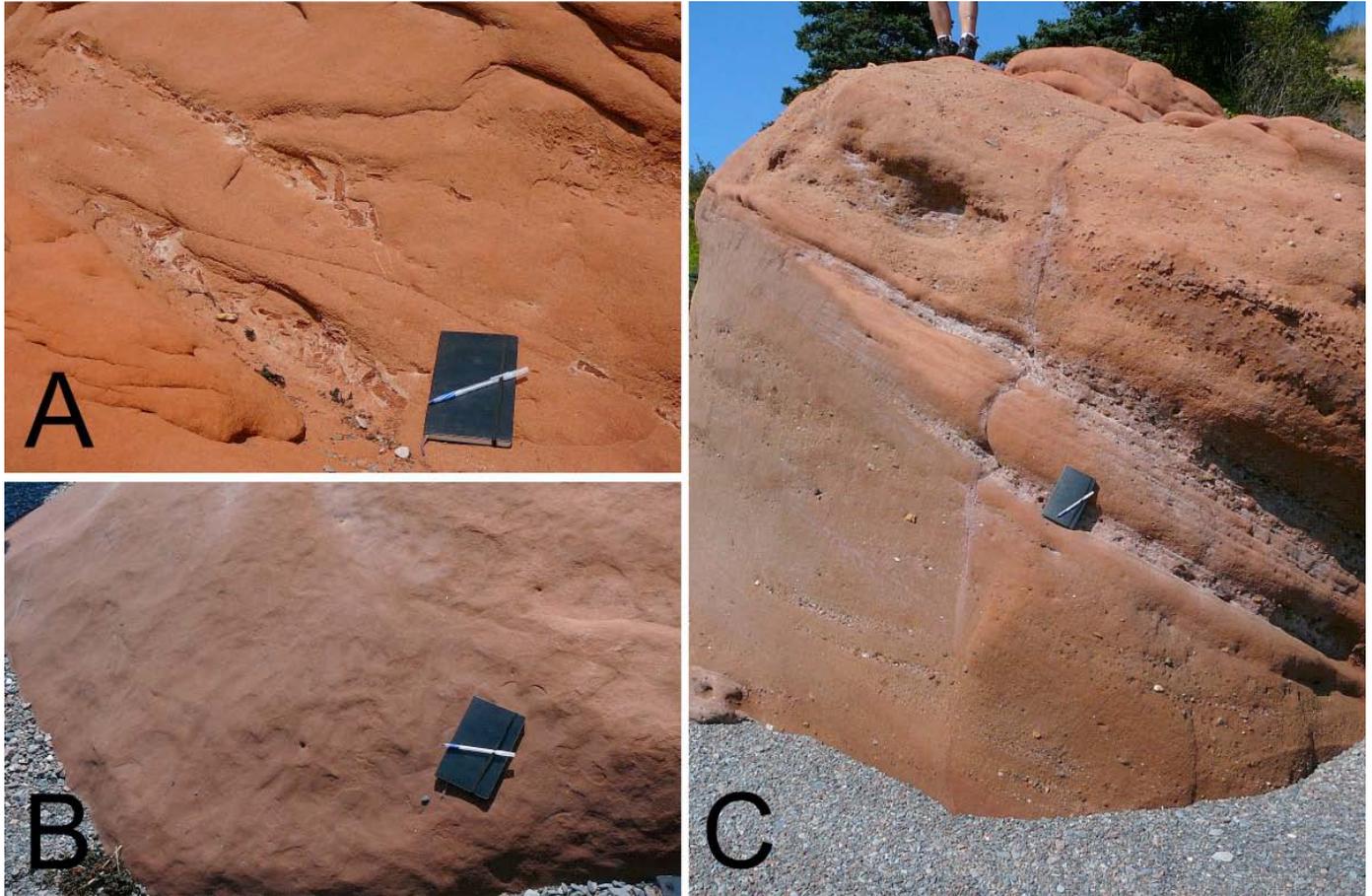


Figure 3. (top) Photographs of the Blomidon formation at Cape Chignecto Park, showing (A) rip-up clasts, (B) bioturbation, and (C) channels.



Figure 4. (right) Photograph of the 'vertical debris flow', showing larger grain sizes cutting vertically through the sedimentary layers.

Moving to the west, we cross a fault strand that cuts onshore, moving from the Blomidon sandstone hanging wall to the Carboniferous-age foot wall. The foot-wall rocks are gray sandstone and mudstone layers, at least one section of which was organic rich. The fault zone deformation is primarily within the foot-wall rock, as indicated by the large amount of strain features (Figure 5). Upon careful inspection, one can also find evidence of hanging wall rocks (red sandstones) in the fault zone. The center of the fault zone is filled with cataclasite bedrock, and the fault gouge zone has fibrous gypsum. Much of the central fault zone is highly eroded (Figure 6).



Figure 5. Strain features in the foot wall at Cape Chignecto showing (A) folding associated with transpression, (B) fault gouge on the right, and lenses of more competent rock that have not been completely ground up on the left, with slickensides showing the shear direction, and (C) a piece of shale that has been severely fractured and mineralized with quartz and gypsum.

The entire fault system accommodated approximately 2000 km of displacement. This fault strand likely had a significant amount of slip, as it has developed a fault zone of nearly 1 km width. The fault system underwent right-lateral slip in the Paleozoic, then left-lateral slip in the Mesozoic. Later in the Mesozoic, it accommodated compressional motion. From their orientation, the observed large folds appear to be from the left-lateral phase of slip. Alternatively, the folds could be associated with Reidel shears in the fault zone.

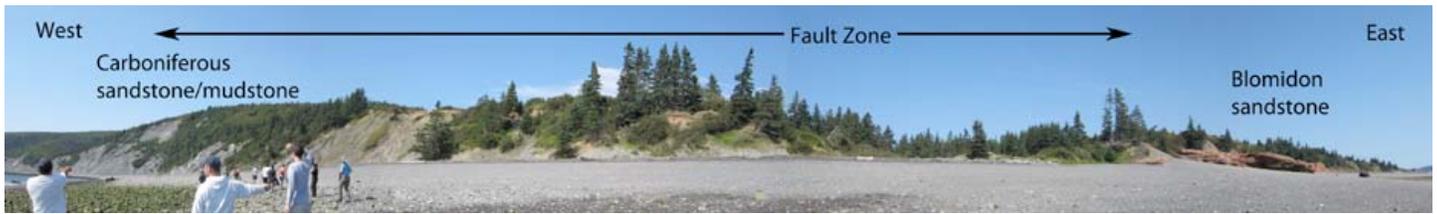


Figure 6. Panorama of Cape Chignecto, facing north, showing the large extent of the fault zone.

Clarke Head

The Cobequid fault segment of the Minas fault zone cuts onshore east of Cape Chignecto at the location of the previous stop, and strikes further inland in an ESE direction (Figure 7). At Clarke Head, the most outboard exposure of the MFZ on the northern Minas shore is found (Waldron et al., 2005). To access the site from Parrsboro, drive east on Two Islands Road approximately 4 km. Park at Glooscap Park Campground, and walk down the beach stairs. Walk east (left) along the beach.

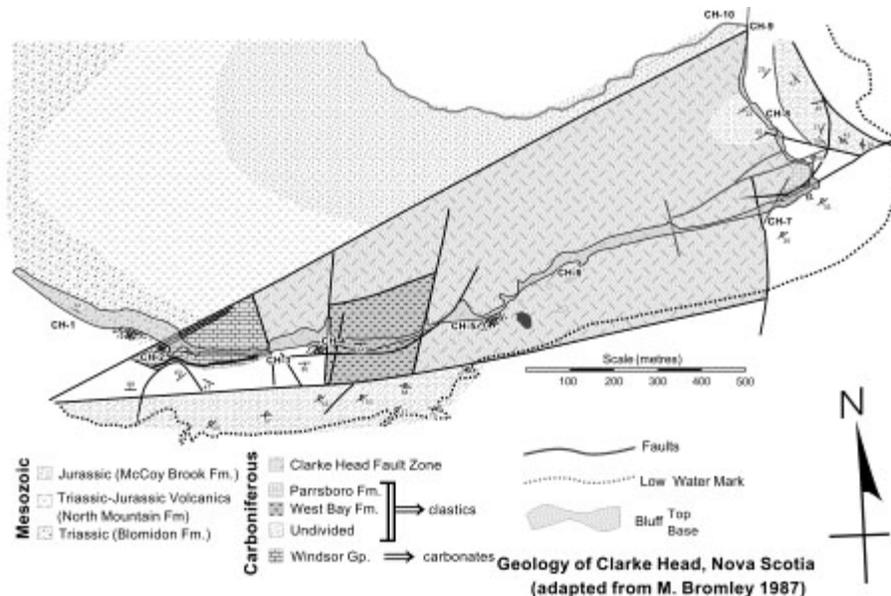


Figure 7. Geology of Clarke Head, with smaller blocks separated by fault strands off the main Cobequid fault (Waldron et al., 2005).

Beginning near the beach stairs, the Blomidon Formation of Triassic red sandstone and siltstone is exposed, dipping shallowly to the northeast. These rocks were deposited in a playa-type environment, with sediments from both lacustrine and hyper arid environments. This is supported by the observed sand-patch texture, cyclical strata, and ripples (Figure 8, following page).

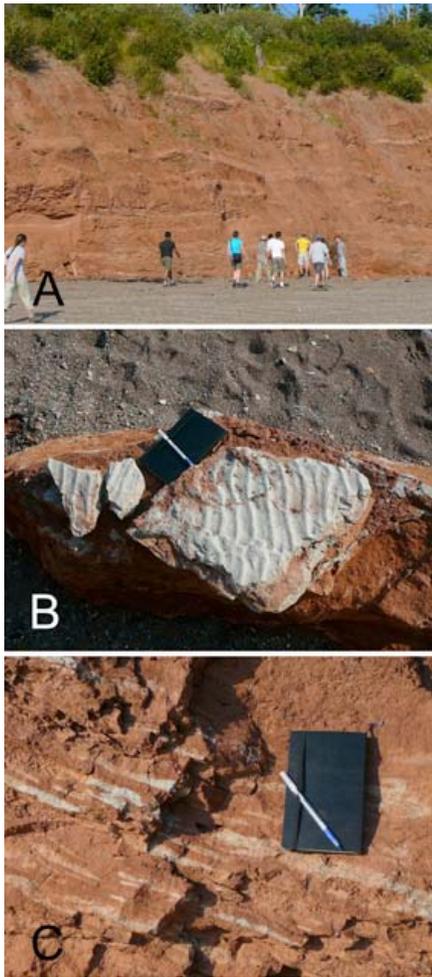


Figure 8. Features of the Blomidon Formation observed at Clarke Head. (A) Cyclic layering, (B) fossil ripples, and (C) sand-patch texture.

Continuing east, the Blomidon strata are overlain by North Mountain Formation basalts. The basalt is Jurassic in age, associated with the onset of rift volcanism. More North Mountain basalt can be seen to the south across the Minas Channel, where Cape Blomidon is visible. At the first sharp bend in the cliff, the Blomidon and North Mountain basalt are juxtaposed next to a section of light gray Windsor Group limestone by a strike-slip fault strand (Figure 9). The Windsor group limestone is Carboniferous in age, and contains *Gigantoproductid* brachiopod fossils and tension gashes.

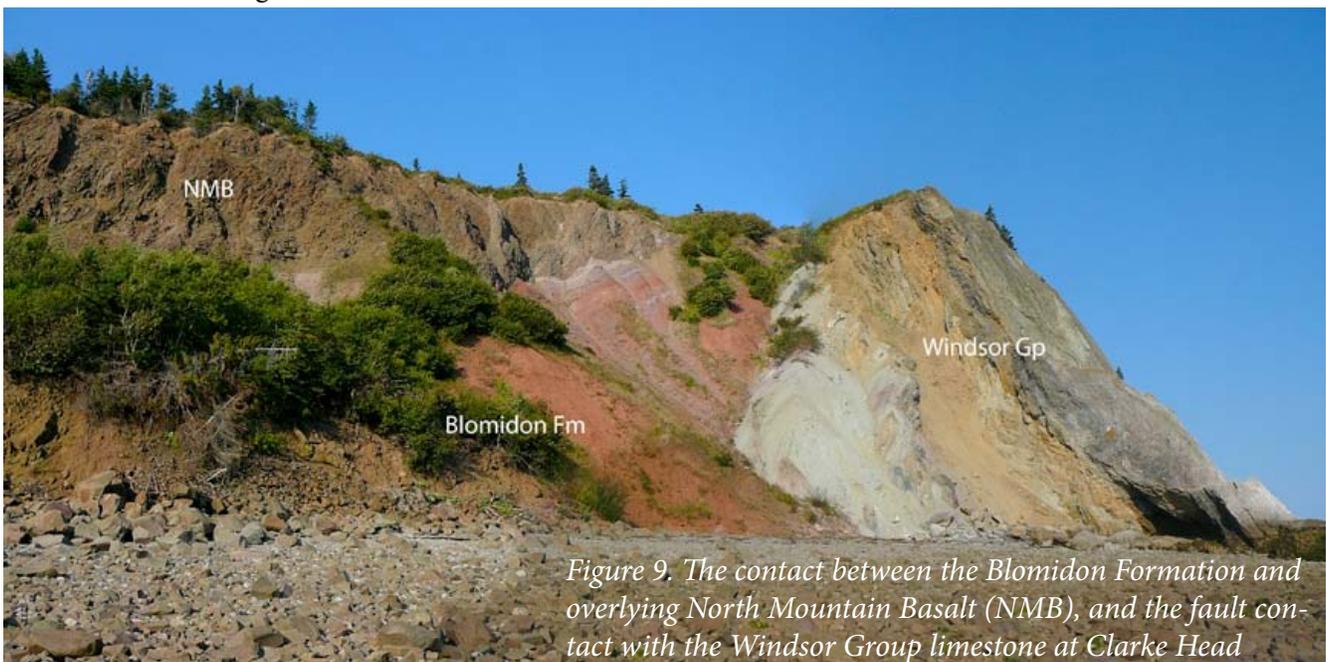


Figure 9. The contact between the Blomidon Formation and overlying North Mountain Basalt (NMB), and the fault contact with the Windsor Group limestone at Clarke Head

Farther east, the limestone becomes less and less coherent, marking progress through the fault zone. Numerous small faults can be observed, as shown in Figure 9. Blocks of garnet-grade schist and other metamorphic rocks originating deep in the Minas fault zone are contained in breccias. This is one of the few locations in southern Nova Scotia where Precambrian basement is exposed. Gypsum becomes common in the breccia matrix. Approximately 0.8 to 1 km from the beach stairs, megabreccia with 1.5 m clasts are observed, held together by gypsum. And finally, a dramatic salt tongue that has been disaggregated can be observed in the fault core (Figure 10). This salt is not the cause of the fault motion, but likely was dragged up or took advantage of the mess the fault made. Continuing east would show a transition out of the fault zone back into more coherent rock. Further information on geologic stops in this area can be found at <http://www.ualberta.ca/~jwaldron/nsfieldtrip/Parrsboro.htm> .



Figure 10. Block of gypsum with mylonites developed from the fault core at Clarke Head.

References

Godfrey, S.J., Fiorillo, A.R., and Carroll, R.L. 1987. A newly discovered skull of the temnospondyl amphibian *Dendrerpeton acadianum* Owen. *Canadian Journal of Earth Science*, v. 24, p. 796-805.

Waldron, John W.F., Joseph Clancy White, Elizabeth MacInnes, and Carlos G. Roselli, 2005. Transpression and transtension along a continental transform fault: Minas Fault Zone, Nova Scotia. Geological Association of Canada, Joint Meeting Field Trip Guide, Chapter 5, Halifax. <http://www.ualberta.ca/~jwaldron/nsfieldtrip/Parrsboro.htm>

(modified from Olsen and Et-Touhami, 2008)

- 7:30** Wake up
- 9:00** Depart Five Islands Campsite
- 9:30** Arrive at Fundy Geological Museum
- 10:15** Depart Fundy Geological Museum
- 10:30** Arrive at Wasson Bluff Protected Area
- 13:40** Finish Wasson Bluff Traverse; Lunch
- 14:15** Depart Wasson Bluff
- 15:00** Arrive at Five Islands Provincial Park Beach Trail
- 18:00** Return to **Five Islands Provincial Park**

Stop 1: Fundy Geological Museum (45° 23.965'N 64° 19.392'W)

The Fundy Geological Museum is located at 162 Two Islands Road in Parrsboro and opens at 9:30 AM. Opened in December 1993, the Fundy Geological Museum in Parrsboro, Nova Scotia attracts over 24,000 visitors year round. The museum includes an exhibition gallery, lab space, a multi-purpose room, gift shop and administration offices. The Fundy Geological Museum mission statement is to be the world centre for experiencing geological history interpreted from the unique features of Nova Scotia's Fundy region. The establishment of the museum was triggered by an announcement in January 1986 by the National Geographic Society that Paul Olsen and Neil H. Shubin of Harvard University had found at Wasson Bluff, near Parrsboro, Nova Scotia, over 100,000 pieces of fossilized bone of ancestral crocodiles, large and small dinosaurs, lizards-relatives, sharks and primitive fishes. It had taken 10 years of determined scouring to uncover this cache of bones, which were fingernail to pencil size. Funded by the National Geographic Society and carried out in co-operation with the Nova Scotia Museum, their find was the biggest in North America that dates from around the Triassic-Jurassic boundary. Not only does the Fundy Geological Museum house the majority of the fossils from Wasson Bluff, but it also has collections from other Triassic-Jurassic sites on the north shore of the Minas physiographic basin, but also Paddy Island and a large collection of Carboniferous vertebrate material from the area.

Stop 2: Wasson Bluff Protected Area Traverse (45° 23.965'N 64° 19.392'W)

To arrive here, drive 8.4 km (~5.2 miles) to the east on Two Islands road. There is an access trail that leads you to the beach. The hike will go ~2km west. There, a small road connects back to Two Islands Road, albeit 2.2 km west of the initial spot. If there are multiple vehicles, then leaving one here to take the drivers back to the start point to return with the rest of the vehicles is a good idea.

The spectacular exposures at Wasson Bluff reveal the interplay between tectonics, sedimentation, and taphonomy described by Olsen et al. (1987, 1989) and Olsen & Schlische (1990). Wasson Bluff is the best place to see evidence for coeval faulting and sedimentation during left-oblique slip of the Minas fault zone (Figure 1, following page). Triassic formations vary markedly in thickness in this area. At Wasson Bluff proper the Blomidon Formation consists of only a veneer of conglomeratic sandstone overlain by finer clastic rocks. The Early Jurassic McCoy Brook Formation fills fault-bounded wedge and trough-shaped basins developed on the faulted upper surface of the North Mountain Basalt. Two well-exposed microbasins (eastern and western) are present here,

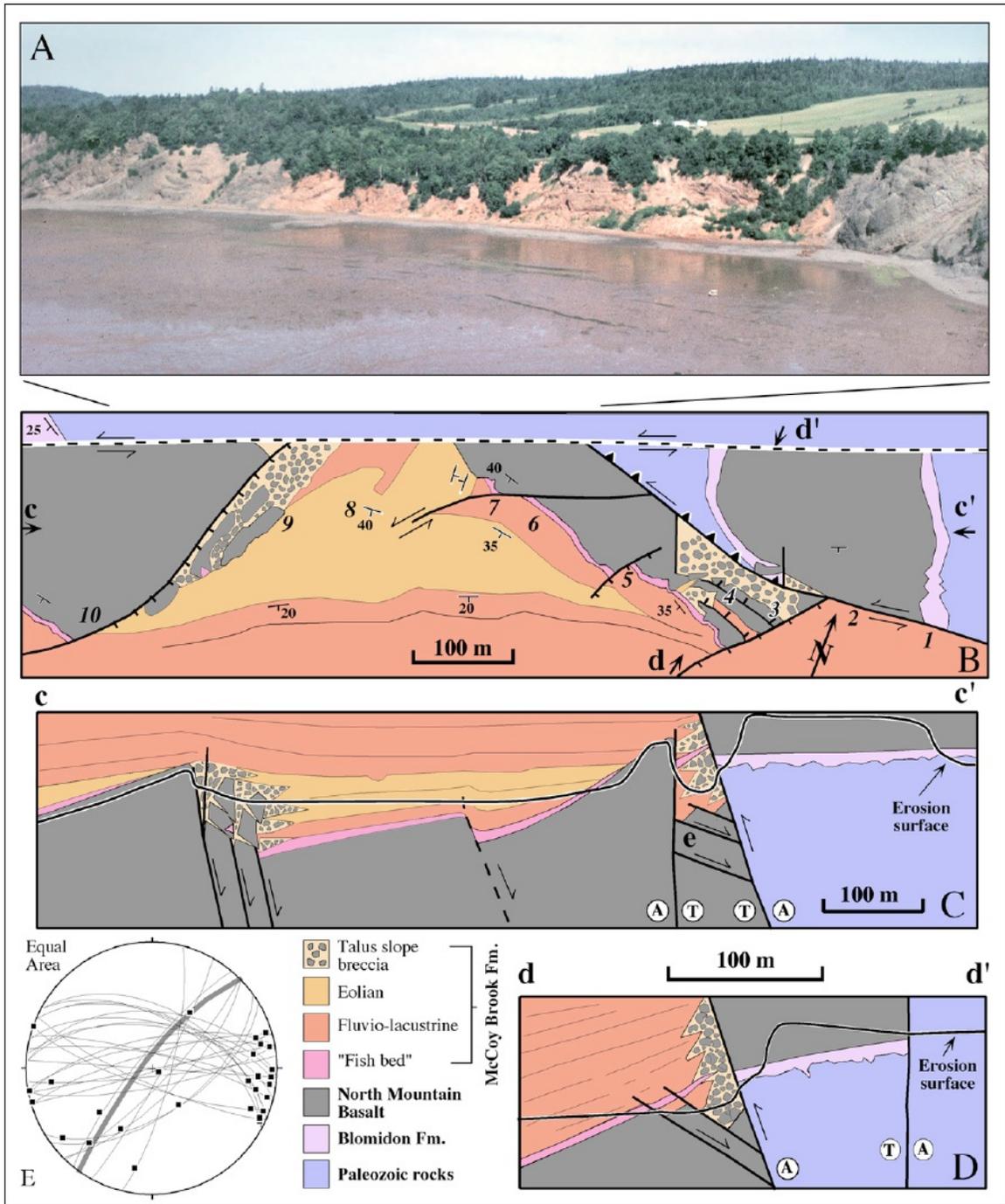


Figure 1. Wason Bluff area, Stop 2.2: (A) helicopter photo of middle segment of the Wason Bluff transect; (B) bedrock geological map of eastern 3/4 of Stop 2.2 transect (1-10 are Stations); (C) cross-section c-c'; (D) cross-section d-d'; (E) equal area lower hemisphere stereographic projection showing attitude of faults (thin great circles), slickenlines (boxes), and average orientation of Neptunian dikes (shaded great circle) on west side of Wason Bluff area, all consistent with northwest-southeast extension. From Olsen & Et-Touhami, 2008.

filled with lacustrine, eolian, and minor fluvial units as well as basalt talus deposits. Paleo-fault talus slope deposits and slide blocks are common in these sub-basins. McCoy Brook Formation debris flows consist exclusively of North Mountain Basalt clasts, testifying to the localized uplift of the lava flows (Tanner & Hubert, 1991). Left-lateral strike-slip and normal faults cut formations of all ages. Neptunian dikes are common in the North Mountain Basalt. The topography generated by active faulting during sedimentation produced a complex relationship between lacustrine fluvial and lava flow units. The high relief and erratic, often very rapid sedimentation related to this relief has resulted in an unusually dense accumulation of fossils.

Station 1: At the immediate eastern end of the exposure and just to the west of the small stream, the unconformity between Blomidon strata and Carboniferous basement is locally exposed in the modern talus slope (Figure 2). The Blomidon Formation is dipping towards the beach (i.e., southward) and consists of well bedded grey, red and purple sandstone and mudstone. The Carboniferous rocks are continental red-beds. The contact between the Partridge Island member (upper Blomidon Fm) and the North Mountain Basalt is faulted (Figure 3). The faults cannot be followed in the basalt. The Partridge Island Mbr is conglomeratic and then fines up to siltstone just beneath the North Mountain Basalt. Hexagonal cooling joints are well developed immediately above the contact. The Blomidon Fm thickens towards the south (Clarks Head).



Figure 2. Outcrop of Carboniferous red-beds at beginning of hike.



Figure 3. Fault that places the North Mountain basalt in contact with the Partridge Island Mbr of the upper Blomidon Fm.

Station 2: The bulk of the North Mountain Basalt from here westward to the eastern sub-basin is mostly brecciated basalt (Figure 4). Olsen et al. (1989) hypothesized that this zone marks a wide, predominantly left-lateral fault zone. Splays of this fault zone trending at a higher angle to the cliff face are present. The brecciation suggests that faulting under near-surface but not exposed conditions; some preserved chloritized slickensides probably formed at greater depth.



Figure 4. Rocks of the North Mountain Basalt brecciated by left lateral fault zone.

Station 3: A fault-bounded block of paleo-fault talus slope deposit marks the eastern end of the eastern sub-basin (Figure 1). The talus deposits are in fault contact with the North Mountain basalt. This sub-basin is easily recognized by the distinctive deposits consisting of angular clast-supported basalt breccia in a matrix of sandstone or mudstone (Figure 5). Large blocks of several basalt flows are common, many of which are themselves broken into smaller blocks with very small amounts of lateral movement. Sediment-filled voids among the basalt clasts show stratification that has yielded consistent bedding orientations throughout the outcrop, indicating deposition of the matrix after the accumulation of the basalt clasts. These deposits at Wasson Bluff are invariably associated with faults and have all the characteristics of talus slope accumulations and abrupt relief (Tanner & Hubert, 1991). These McCoy Brook talus slopes contained many large empty spaces in which small animals could live. Unfortunately, most bones and skeletons are truncated at one end or another by small faults, and much material appears ‘chewed’. Bone-bearing coprolites occasionally occur, suggesting that at least some of the remains may have been dragged into the talus piles by a predator or scavenger. Tetrapod bones may be more abundant in this talus slope breccia than anywhere else in eastern North America.



Figure 5. Paleo-talus sediments exposed. The basalts were eroded from an adjacent cliff. The matrix is formed by sediments that progressively filled in the voids between blocks. The picture on the right shows a diffuse stratification, showing it was a gradual infill process.

Station 4: A high-angle reverse fault separated the talus slope deposits from the vertical Carboniferous basement rocks to the immediate northwest, which are unconformably overlain by the Blomidon Formation, in turn overlain by North Mountain Basalt (Figure 1, 6). This is an extension of the same contact surface seen at Station 1. South of the talus slope deposits are a series of north-dipping normal fault-bound “dominoes” of North Mountain Basalt (Figure 1). Basins that developed in these fault blocks are filled with McCoy Brook Formation, which consists of brown, at least partially eolian, sandstone. The southernmost outcrop of basalt adjacent to the “dominoes” is overlain by the Scots Bay Member. Note that if the Scots Bay beds are rotated to horizontal, the reverse fault adjacent to the talus slope deposits becomes a normal fault, and the normal faults associated with the “dominoes” are antithetic to it. Hence, the sandstones and overlying talus-slope deposits probably accumulated on the downthrown side of a normal fault, creating a “buttress unconformity” (deposition against a steep slope, in this case a fault plane). The antithetic faults are characterized by basalt breccias that appear to be filling surficial fissures, suggesting that this was very near to the surface at the time of faulting.



Figure 6. View of the reverse fault which places Carboniferous rocks over the Blomidon Fm (dark red layers under the North Mountain basalt in the upper right corner). The normal fault on the left side places the basalt in contact with the Blomidon Fm through a buttress unconformity.

Station 5: The top of one flow, another thin flow, and a third upper flow of the North Mountain Basalt are visible in the next stretch of outcrops. Here, the basalt is mostly massive and disturbed only by small faults. The green, red, and gray flows are each separated by vesicular horizons. Locally there are clasts of sedimentary rocks between the flows as well (Figure 7). There are also zones of alteration that follow the contacts between flows. This could have been due to fluid flows that occurred after deposition of the basalts. The existence of a diffuse layering in sediments filling the porosity in the altered areas supports this interpretation. This particular stretch of outcrop also contains numerous neptunian dikes (Schlische & Ackermann, 1995), formed by the active extension of the North Mountain Basalt and filled with sediment from above.



Figure 7. Contact between successive basalt flows marked by clast of sedimentary rocks. Larger reddish layers represent Neptunian dikes.

Station 6: The basalt flows in the region dip to the southwest. The thick uppermost gray basalt flow is overlain by the Scots Bay Member, which can be excavated in the beach. The Scots Bay Member and other strata of the McCoy Brook Formation strike directly into the North Mountain Basalt in the cliff face, along which an east-west striking fault runs (Figure 8). The fault has offset the North Mountain Basalt—Scots Bay Member contact by approximately 25 m (Figure 1). The fault zone is marked by a mineralized, slickensided plane with subhorizontal slickenlines. The basalt on either side of this plane is clearly brecciated. The upper surface of the basalt contains numerous neptunian dikes, the offset of which shows the motion to be almost pure strike-slip, agreeing with the slickenline directions.



Figure 8. Evidence for faulting. Picture on the right shows that layering runs right into the North Mountain basalt, indicating that there must be a fault between them. The picture on the left shows sub-horizontal slickensides, indicative of strike-slip motion on the fault.

Station 7: The eastern end of the western sub-basin is marked by the onlap of the Scots Bay Member onto the upper surface of the North Mountain Basalt (Figure 1, 9). Onlap of the Scots Bay Member onto the basalt is marked by a gravely mudstone very rich in disarticulated fish and small tetrapod bones, the “fishbed”. The upper portions of the lacustrine sequence pass upward into red mudstone and sandstone beds. This unit dips steeply at the outcrop of the Scots Bay Member but it flattens out and reappears in the beach a few meters to the west. The mudstone has rhizolites which indicates that there are paleosols here.



Figure 9. The McCoy Brook Fm onlapping onto the North Mt basalt. The lower-most layer of the McCoy Brook here is the “fish-bed”.

Station 8: This area lies near the center of the triangular western sub-basin and the cliffs (Figure 1) are representative of most of its fill. The fill lies stratigraphically above the sediments seen at the previous stop. Most of this fill consists of a NW- to SW dipping largely eolian dune sandstone. According to Hubert & Mertz (1984), at least 48 m of dune sand are present in this basin. Toward the western end of the sub-basin, basalt clasts within the dune sands become more abundant and larger and at the western end of the sandstone outcrops (Figure 1), eolian dune sands abut against a large slide block or domino of relatively intact basalt, producing an 8-m-high paleocliff with adjacent talus cones (Hubert & Mertz, 1984) (Figure 10). Successive first order surfaces in the eolian sandstone are veneered with basalt boulders (some larger than 1 m) that rolled down-slope and were buried by advancing dunes and ridges. As the northwestern normal fault boundary of the sub-basin is approached, eolian dune sands interfinger with paleotalus slope deposits and another slide block of tectonically-disrupted basalt. These relationships indicate that the faulting responsible for sub-basin formation and sedimentation were coeval.



Figure 10. Picture showing the faulted contact between the eolian sandstones on the far right and the basalts on the left. There are 2 blocks that are “falling” towards the sandstones. The one most to the right is probably a slide block, while the one in the center of the picture is probably a rider block.

Station 9: A basalt-clast fault-related breccia with abundant sediment matrix crops out at the contact between the upper North Mountain Basalt and sedimentary strata to the west. Its bedding attitude is obscure. It is bound on its southwest by a high-angle fault striking WNW with subhorizontal slickenlines. Immediately to the west of this fault, strata within the McCoy Brook Formation, including thinly-bedded purple-gray-brown lacustrine units, have vertical bedding (Figures 1, 11). This region of upturned strata is due to later Jurassic compression during inversion (Withjack et al., 1995). The meter-scale laminated and thin-bedded vertically oriented purple-gray-brown interval appears to be a lacustrine interval above the Scots Bay Member.

Figure 11. Picture showing the contact between the North Mt basalt on the right and the sub-vertical layers of the McCoy Brook Fm to the left. A diagrammatic section is shown in Figure 13.



Station 10: As we walk to the west, dips in the McCoy Brook Formation quickly shallow. The remainder of the outcrop consists of debris flows of the McCoy Brook Formation, and interbedded and overlying red sediments (Figure 12). The clasts of the matrix-supported flows consist of North Mountain Basalt. Individual debris flows are 10's of m thick. Several WNW-striking, left-lateral(?) faults cut these debris flows. The interbeds of red clastics are paleosols riddled with abundant root mottles and purplish carbonate nodules. Only a thin section of sedimentary strata overly the debris flows, but these appear to be normal McCoy Brook sand patch mudstones with gypsum nodules. A view to the west and southwest (towards Clarke Head) shows a reverse fault throwing North Mountain Basalt over McCoy Brook Formation which is now dipping to the NE. The synclinal shape of the overall structure is supported by measurements of McCoy Brook bedding attitudes in the beach (Figure 13). This synclinal structure bears a striking resemblance to structures seen in seismic lines along the projection of the Minas fault zone into the Bay of Fundy (Withjack et al., 1995; Wade et al., 1996). It is worth noting that the thickness of the Blomidon at Clarke Head is in excess of 100 m, showing a rapid increase in thickness from the Wasson Bluff area, presumably due to Triassic faulting and subsidence.

This way of leaving is approximately 2.8 km south west of where we parked.



Figure 12. Picture showing a faulted layer of debris flow characteristic of the upper Mc Coy Brook Fm in the synclinal trough between Wasson Bluffs and Clarke Head.

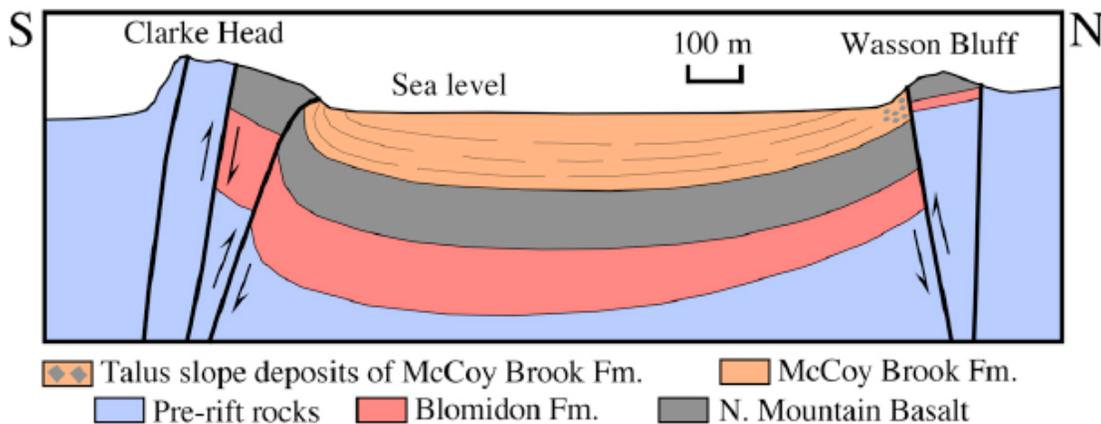


Figure 13. Cross-section from Wasson Bluff to Clarke Head. Interpretation is from Withjack et al. (1995) in which the stratigraphy is strongly modified by syndepositional tectonics and there is a thick post-basalt unit (McCoy Brook Fm.). From Olsen & Et-Touhami, 2008.

Stop 3: Five Islands Provincial Park (45° 23.671'N 64° 3.710'W to 45° 23.005'N 64° 02.211'W)

Directions from Stop 2:

Go west on 2 Island Rd 65 km

Turn right at Main St 0.2 km

Slight right at Eastern Ave (Hwy 2) 1.5 km

Continue on Hwy 2 24.6 km

Turn right at Bentley Branch Rd (access to Five Islands Provincial Park) 3.5 km

Turn right towards beach 0.5 km

Enter Five Islands Provincial Park from the main entrance and proceed to the beach parking area. The structures exposed at Five Islands Provincial Park are associated with the east-trending Minas fault zone, the northern boundary of the Minas subbasin (Figure 14). The Minas fault zone has had a complex tectonic history, and the sections are much more structurally complex. This section has a structurally shortened transect across the Triassic- Jurassic transition with thick sections of the Triassic-Jurassic Blomidon Formation and North Mountain Basalt and McCoy Brook Formation beautifully displayed. During Palaeozoic time during accretion of allochthonous terrains, it was a convergent, right-lateral strike-slip fault zone. During early Mesozoic time, it became a divergent, left-lateral, strike-slip fault zone. During middle Mesozoic time, the Minas fault zone again became a convergent, right-lateral, strike-slip fault zone, allowing considerable shortening of the hanging wall. Structures visible along this traverse are consequences of both the early Mesozoic extension and middle Mesozoic contraction.

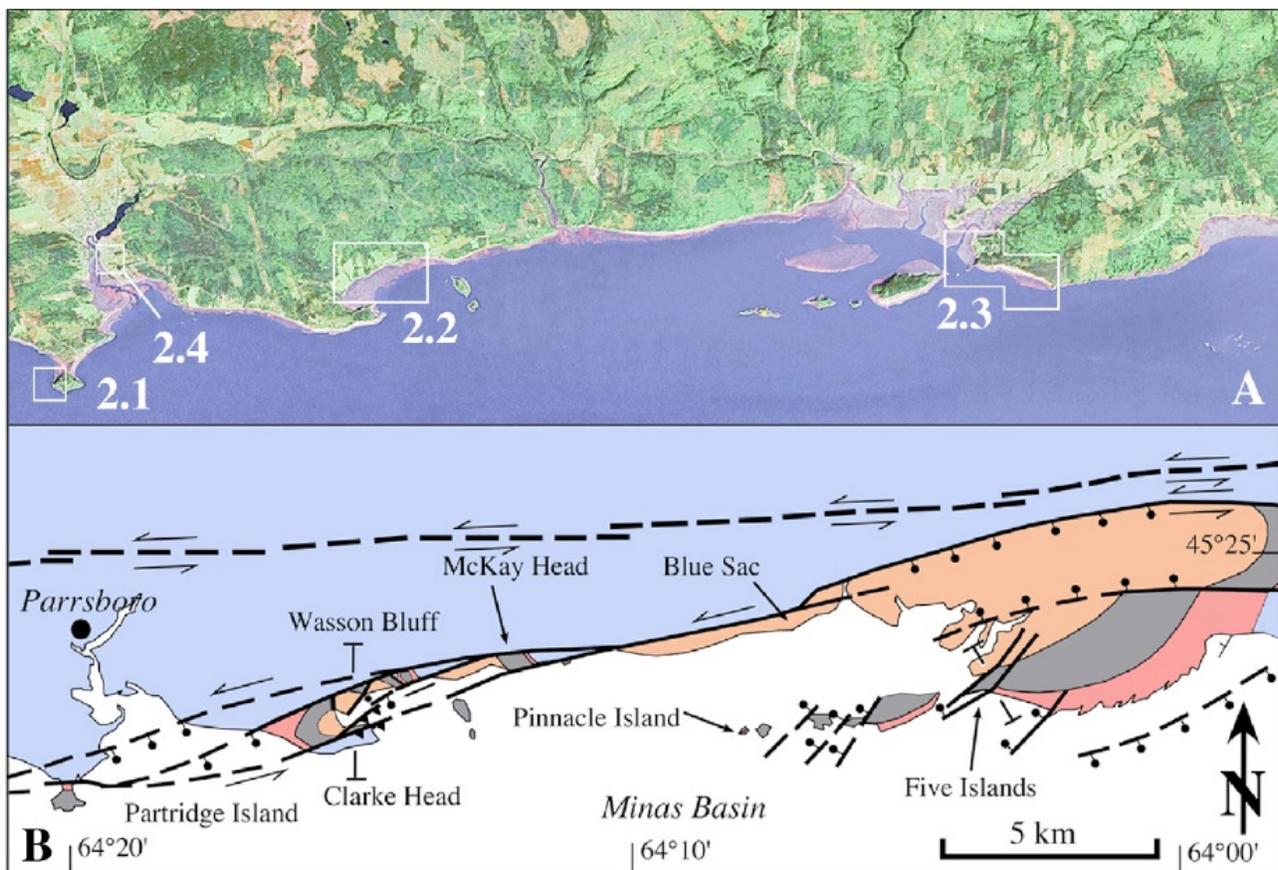


Figure 14. Index map for stops along the north shore of the Minas physiographic basin. (A) Satellite photograph of field area with field stops shown as white boxes 2.1 – 2.4 . Stops 2.4, 2.2 and 2.3 correspond to stops 1, 2, and 3 respectively of our trip. (B) Detailed geological map of field area showing areas discussed in text and positions of cross sections shown in Figure 1 and Figure 15. Position of map in Figure 1 is white box next to 2.2. From Olsen & Et-Touhami, 2008.

Walk down to the beach and continue southeast about 0.6 km along the cliffs to the basalt peninsula of Old Wife Point. Climb up onto the peninsula itself; facing the cliffs from the Old Wife, we have a great view of one of the most spectacular outcrops of Triassic-Jurassic strata in eastern North America (Figure 15). Then proceed east-southeast along the beach about 2 km and round the distinctive point of brown sandstone and stop inside the small sandstone cove. This is Station 1 and from here we will proceed backward along the direction we came stopping at the other stations.

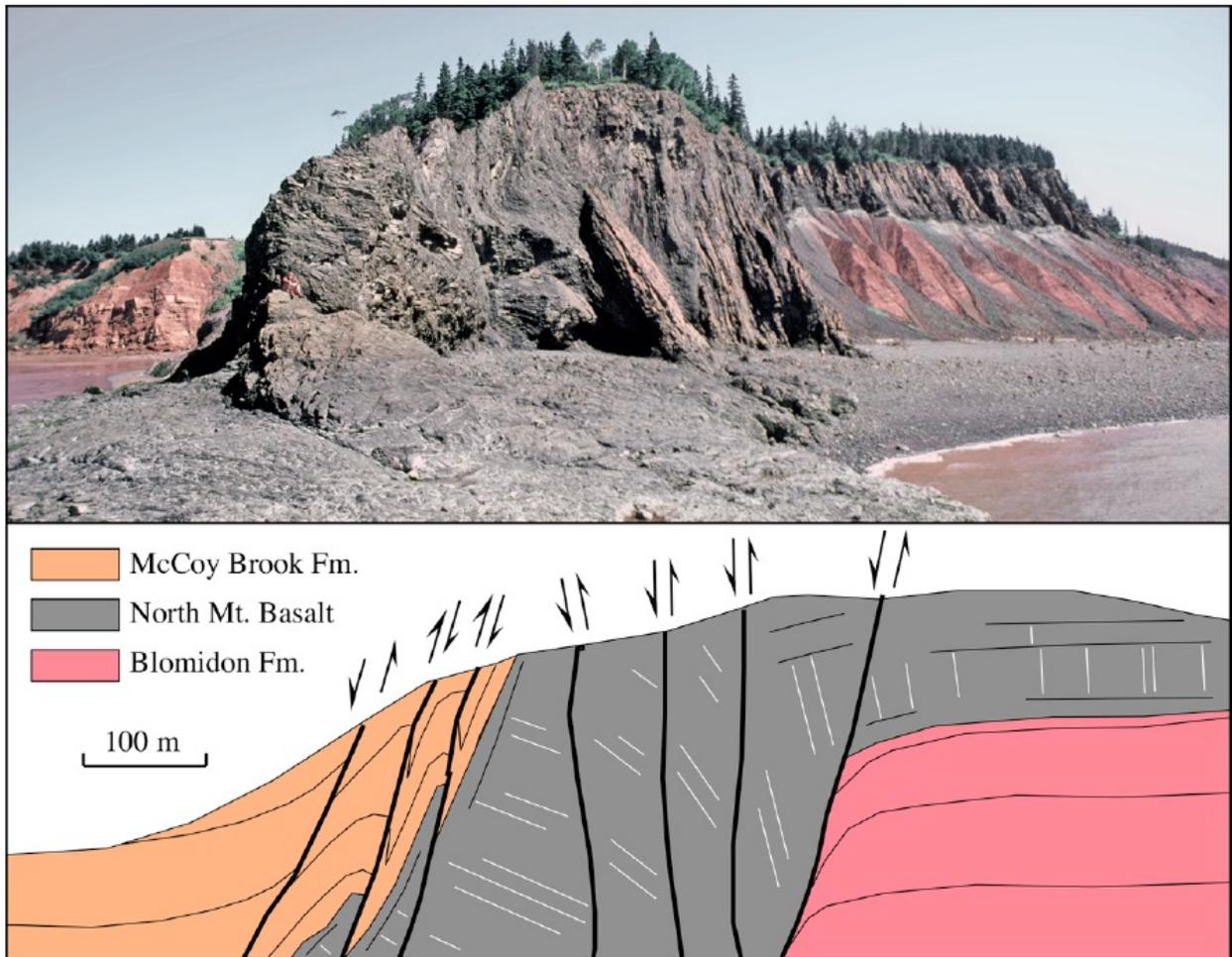


Figure 15. View from Station 3, Stop 3: above, wide-angle view to the NE from the Old Wife showing Blomidon Formation conformably overlain by North Mountain Basalt on the right, down faulted North Mountain Basalt straight ahead; and down faulted and reverse faulted McCoy Brook Formation on left. Below, cross-section (from Withjack et al., 1995). From Olsen & Et-Touhami, 2008.

Station 1. Eolian and fluvial sandstones: These outcrops of the Red Head member of the lower Blomidon Formation consist of about 33 m of eolian dune sands and several meters of interbedded fluvial sandstone and pebbly sandstone and conglomerate (Hubert & Mertz, 1980, 1984) (Figure 16). A southwest striking, northwest dipping fault separates the Red Head member from the middle Whitewater member of the Blomidon Formation (Figure 16). Strata immediately to the west of the fault are comprised of a distinctive suite of deformed, mostly lacustrine beds that correlate to the middle of the Blomidon Formation. The upper strata is brecciated downward to contorted sands and siltstones, and overlain by a granule-bearing conformable sandstone bed. Lower strata are deformed by numerous bed-specific normal faults of various sizes (Figure 17). This deformation was interpreted by Olsen et al. (1989) and Ackermann et al. (1995) as a result of salt (halite) dissolution. Tanner (2003), however,

interpreted the deformation as due to a strong synsedimentary seismic event, namely the impact of the bolide that produced the Manicouagan structure. Olsen argues that a seismic event was not responsible because the structures and the internal stratigraphy require multiple dissolution and deposition events and a downward, not upward motion of material. Furthermore, the age of the strata is 210 Ma, significantly younger than the 215.5 Ma date for the impact structure (Ramezani et al., 2005). Proceed west-northwest about 550 m.



Figure 16. Eolian strata of the Red Head member of the Blomidon Formation (on the right) in fault contact with the lacustrine strata typical of the Whitewater of the Blomidon Fm. Fault is located where the tress come down the cliff. Cyclic layering in the Whitewater Mbr represent Van Houten cycles.



Figure 17. Penecontemporaneous deformation attributed to evaporite (halite) dissolution SC, marks the level of salt collapse beds. Domino-style faulting at lower right is also attributed to salt dissolution. From Olsen & Et-Touhami, 2008.

Station 2: Sand patch cycles in upper White Water Member: As seen here the rest of the Whitewater member at these outcrops is comprised of vertically varying Van Houten cycles of the sand patch variety (as defined by Olsen (1986), a Van Houten cycle records a complete cycle of the lake-level rise, high-stand and fall) (Figure 16). Hundreds of mostly small-displacement normal faults subtly cut the section (Figure 18), and at least one large northwest dipping fault cuts out enough section to prevent us from measuring a complete section from the basalt through the salt dissolution structures. At the top of the Blomidon Formation the sand patch cycles become indistinct and increasingly mud-rich followed by a transition upward to white weathering mudstones within a few meters of the North Mountain Basalt (Figure 19). This is the Partridge Island member marking the base of TS-IV. The apparent lateral changes in thickness of the white layer and its occasional apparent disappearance are actually the result of the numerous small faults, some of which strike parallel to the face of the cliff. Along the top of the cliff is the lower part of the massive lower flow of the North Mountain Basalt. The entablature of the flow has a thickly splintery fracture, not a hexagonal columnar jointing. This style of jointing persists along the entire length of outcrop of this flow. Proceed west-northwest about 1.3 km back to Old Wife Point.



Figure 18. Pervasive small scale normal faults present in the Whitewater Mbr of the Blomidon Fm.

Figure 19. Picture showing the white layer prevalent at the contact between the Blomidon Fm and the North Mt basalt. Layer is faulted out where not present.



Station 3: Old Wife Point looking north and south: The complex structure of this section of outcrops is described and interpreted in detail by Withjack et al. (1995). Looking to the Northeast, the lower flow of the North Mountain Basalt is truncated by a series of high-angle faults that drop the section down to the west. With each more western fault, the splintery basalt columns are rotated progressively counter clockwise (Figure 15). Further to the west are fault slices of vesicular to massive and columnar basalt that appear to be parts of the upper flows of the basalt. Bedding generally is steep to the northwest, but shallows progressively in more westerly fault slices. Looking to the southwest we see the Old Wife sea stack of tectonically brecciated North Mountain Basalt. These faults also dismember the islands that are seen just off-shore of the peninsula. This fault zone marks the transition from Triassic Blomidon Fm to Jurassic McCoy Brooks Fm. Proceed northwest onto the beach and walk along the beach contact with the North Mountain Basalt.

Station 4: Contact with the McCoy Brook Formation: The contact between the McCoy Brook Formation and North Mountain Basalt on the shoreline is partly sheared (Figure 20). At the contact, the McCoy Brook Formation contains polymict cobbles and pebbles of North Mountain Basalt. The northwest steeply dipping red-bed section is sheared and the sense of motion is down to the northwest, consistent with early Mesozoic extension. The McCoy Brook Formation is in fault contact with North Mountain Basalt again on the west. Because the fault is vertical or dipping to the northwest there is older (North Mountain Basalt) over younger (McCoy Brook Formation) and the fault thus has apparent reverse throw (Figure 20). The reverse motion was younger than all the units present here and was probably of middle Mesozoic age (late Early Jurassic to Middle or Late Jurassic). Careful examination of the steep western slope of the North Mountain Basalt reveals several other similar faults and contacts with apparent reverse motion. The most westerly block of basalt is overlain by steeply dipping McCoy Brook Formation red beds, which are again sheared. But they are intact enough to reveal that the contact with the North Mountain Basalt is a normal sedimentary one, albeit somewhat sheared. As we look further to the northwest along the cliff face, the bedding attitudes of the McCoy Brook Formation shallow dramatically and the shearing disappears. A steep cliff face of relatively undisturbed McCoy Brook Formation then continues to the northwest. The reverse faults that indicate later shortening along extensional fault zones is some of the primary evidence for post earliest Jurassic tectonic inversion in the Fundy basin (e.g., Withjack et al., 1995; Schlische et al., 2003). Proceed northwest, along the long cliff outcrops of the shallow-dipping McCoy Brook Formation.

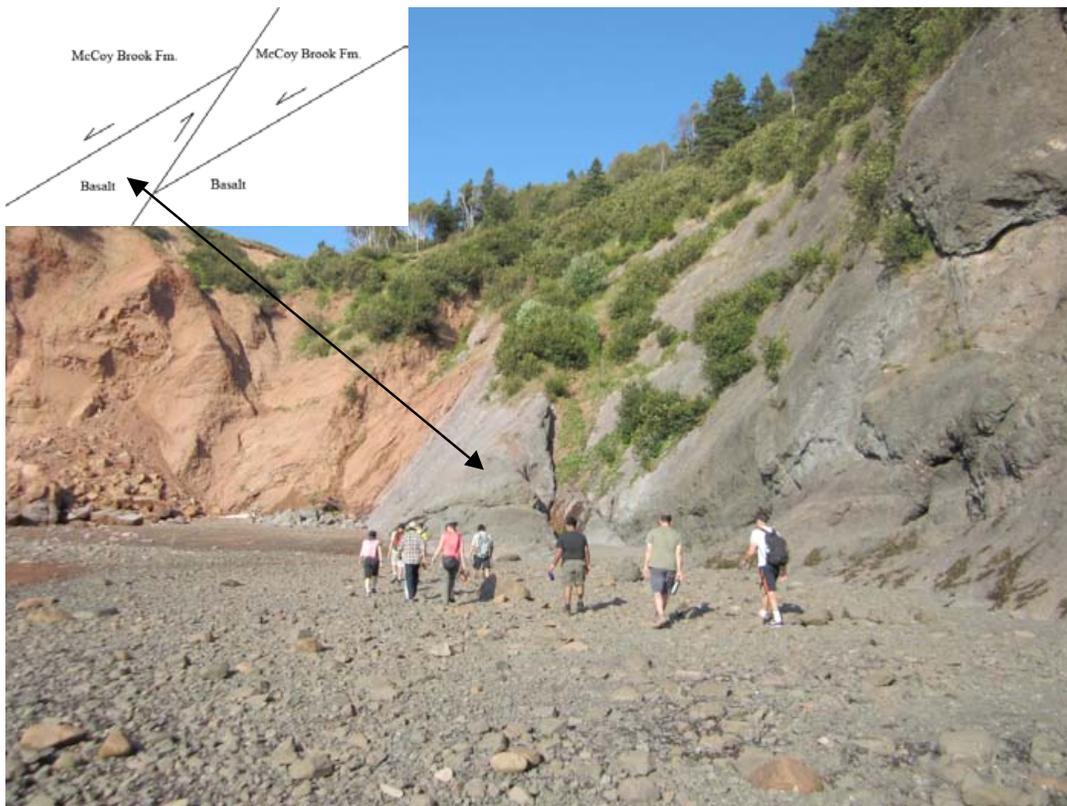


Figure 20. Picture showing the faulted contact between the North Mt basalt and the McCoy Brook Fm. The sedimentary rocks have been down-dropped by a normal fault which dips towards the left of the picture. This fault was later cut by a small reverse fault which uplifted a wedge of basalt over the McCoy Brook Fm. A schematic of this is illustrated in the upper right corner.

Station 5: Cliffs and beach outcrops of the McCoy Brook Formation: There are four main features to be observed in this stretch of cliffs:

Station 5.1: Large-scale channel and delta sequence. This section has thin-bedded mudstones and sandstones succeeded upward by a large sandstone complex part of which has downward tapering tilted surfaces compatible with small delta forests.

Station 5.2: Thin- to thick-bedded, climbing-ripple cross-laminated sandstone alternating with red fine mudstone and claystone with widely-spaced desiccation cracks and reptile footprints that post date the end-Triassic extinction

Station 5.3: Channel-fill and lacustrine sandstones with fish bone- and coprolite-bearing intraformational conglomerate at base.

Station 5.4: Poorly-exposed lacustrine strata of McCoy Brook Formation, with sandpatch cycles and gypsum nodules.

The base of the Blomidon Formation is well exposed in the tidal flat and accessible at low tide and rests unconformably upon Wolfville Formation sandstones and gravels. The basal Blomidon strikes about 270° to 300° and dips about 35° to 40° N, while the Wolfville Formation strikes about 340° to 360° and dips about 20°E. This is probably the best area to see the TS II – TS III unconformity in the Fundy basin. The high degree of discordance implies a long hiatus in this area, which is in agreement with the 14 m.y. gap inferred from the combined bio- and magnetostratigraphic correlation of the Fundy section to the Newark basin time scale.

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Day 6 (August 28th, 2012): The Ordovician-Silurian Arisaig Group

by Jesse Farmer

- 6:30** Wake up, breakfast
- 8:00** Depart Five Islands Campsite to Arisaig, NS
- 10:15** Arrive at Arisaig Point and meet with Professor Mike Melchin, St. Francis Xavier University
- 11:10** Stop 1: Hike eastward from Arisaig Point
- 11:40** Outcrops of Upper & Lower Ross Brook, Beechill Cove, and Bears Brook Volcanics Fms
- 13:00** Lunch at Arisaig Point
- 14:00** Arrive at Stop 2: Arisaig Provincial Park
- 15:30** Depart Arisaig Provincial Park and head west
- 15:45** Arrive at Stonehouse Formation outcrops (John Joe Lane)
- 16:45** Depart for Whycomomagh, NS
- 17:00** Stop for provisions outside Antigonish, NS
- 20:30** Arrive at **Whycomomagh Provincial Park** campsite
- 20:50** Dinner at Whycomomagh restaurant

Stop 1: Arisaig Point and lower Arisaig Group (45.762°N 62.170°W)

Bimodal rhyolites and basalts of late Ordovician age (~560 Ma) are exposed along the northern tip of Arisaig point and along the beach to the east (Figure 1a). Basaltic flows are interrupted by varying degrees of soil formation, with laterally discontinuous paleosols apparent in most basaltic outcrops (Figure 1b). Previous estimates of atmospheric CO₂ from these paleosols suggest Ordovician atmospheric CO₂ levels on order of 8x to 20x present values (Feakes et al., 1989). However, a recent analysis of the chemistry of paleosols east of Arisaig Point suggested warm and humid conditions with alkaline soil pH, a condition lacking a modern analogue and suggestive of little to no vegetation cover during the Ordovician (Jutras et al., 2009). A mafic to felsic transition in the rhyolites of Arisaig Point is attributed to a transition from monsoonal to semi-arid climate.

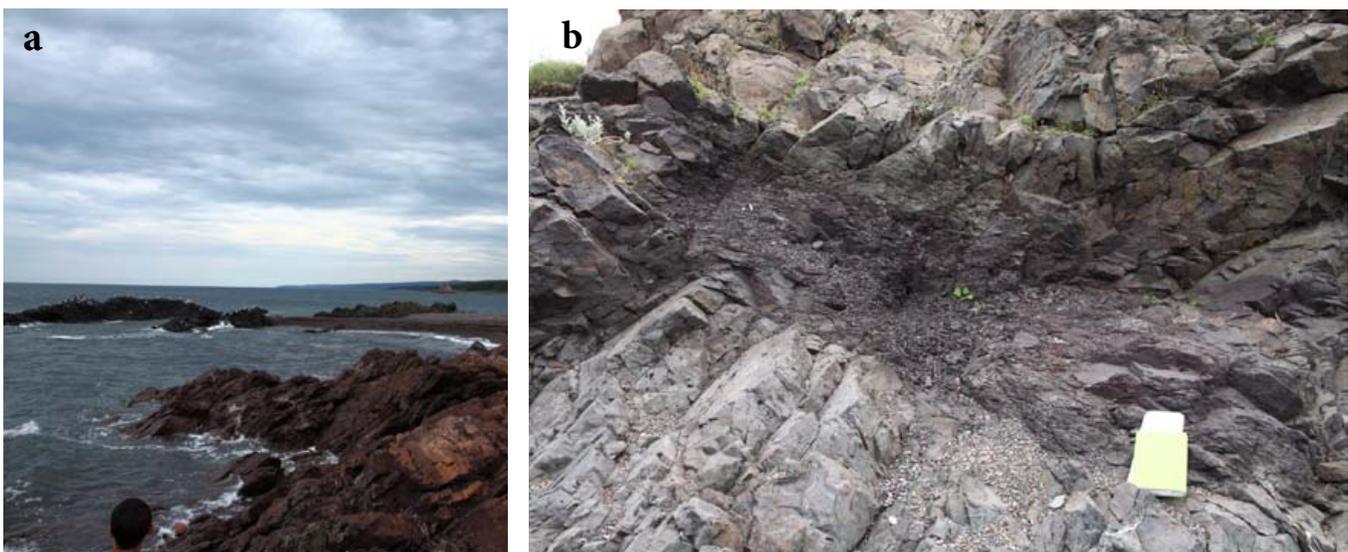


Figure 1. (a) Bimodal volcanics at Arisaig Point; rhyolites in foreground; basalts at left in background.
(b) Paleosol layers developed between basaltic flows.

Approximately 500 m to the east of Arisaig Point, outcrops of the Middle and Lower Ross Brook Formation are exposed immediately along the shoreline, dipping steeply to the WNW (~80°). The Lower Ross Brook Fm is represented by laminated black to gray shales and is particularly rich in graptolites and bentonites. Immediately beneath the Lower Ross Brook Formation (to the W) is the Beechill Cove Formation, represented by a sequence of massively bedded siltstones with coarser, sandstone layers interpreted to be tempestites. The Beechill Cove Fm. is interpreted as representing a shallow, nearshore depositional environment of latest Silurian or early Devonian Age.

To the west, volcanic rhyolites of the Bears Brook Group are exposed along a competent point feature approximately 800 m east of Arisaig Point (Figure 2). Although the contact between the Beechill Cove Fm and the Bears Brook Group has generally been considered unconformable, we examined a newly uncovered exposure along the shoreline that showed evidence for a conformable contact and some evidence of either paleosol formation or intrusion of the Bears Brook Group into the Beechill Cove Fm (Figure 3). The lowermost Beechill Cove Fm appears as greenish-gray, thinly bedded siltstones transitioning to more massively bedded grayish-red siltstones.

Figure 2. Volcanic rhyolites to the east of Arisaig Point.

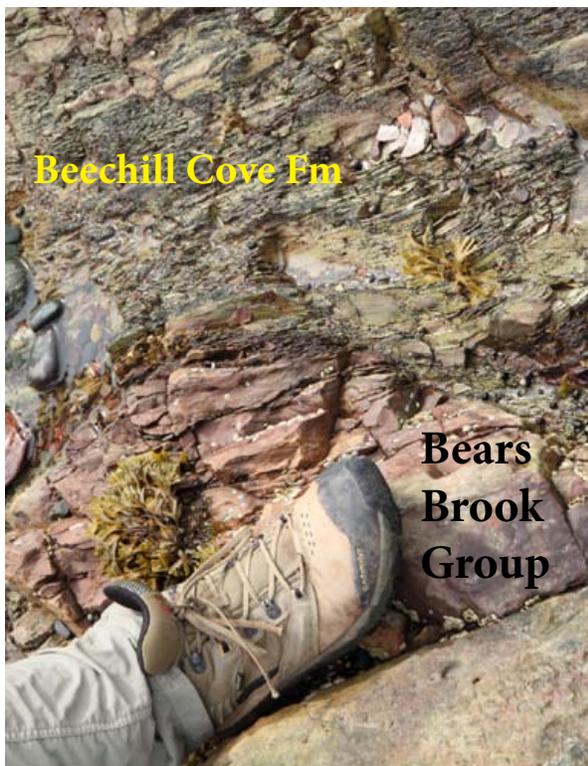


Figure 3. Newly exposed contact between Beechill Cove Formation (top) and Bears Brook Volcanics (bottom) showing conformable contact with finely laminated, discolored layers between the two formations.

Stop 2: Arisaig Provincial Park(45.754°N 62.171°W)

To access Arisaig Provincial Park, drive out of Arisaig Point, turn right onto Highway 245 and head 500 m west. The turn off is on the right. Stay to the left in the campground and there is a small parking area with access to trails and the beach. The Middle and Upper units of the Ross Brook Formation are exposed along the coast to the east of the beach access (Figure 4). Facies show common brachiopod, trilobite, and crinoid fossils that are particularly abundant in massively bedded siltstone layers (interpreted as storm bed deposits, Figure 5). To the west of the beach access and up to Arisaig Brook, foliated shales are exposed that indicate metamorphism. The foliation direction obscures original bedding planes and fossils are nearly absent from these facies.



Figure 4. Fossil hunting in the Middle Ross Brook Formation at Arisaig Provincial Park



Figure 5. Looking perpendicular to the bedding surface at abundant crinoid fossils in a tempestite. Crinoid fossils are oriented both along and perpendicular to bedding.

Stop 3: Stonehouse Formation and Avalon-Meguma collision (45.735°N, 62.211°W)

Along the coastline north of John Joe Lane (approximately 4km west of Arisaig Point along Highway 245), units of bioturbated mudstones and shales with interbedded siltstones and shelly limestones are exposed to the east of the cottage beach access (Figure 6). These units are classified as the Stonehouse Fm and interpreted to represent a shallow marine depositional environment. Above these units, a sequence of red-beds (~10 m) is unconformably overlain by Upper Devonian basalts, with evident faulting separating the units. The competent basalt units form two points to the east of the cottage beach access and one to the right. The rightmost point shows an apparent angular unconformity that is generally coeval with the timing of the collision between the Avalon terrane (on which the Arisaig group was deposited) and the Meguma terrane.



Figure 6. The Stonehouse Formation west of Arisaig Point

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Day 7 (August 29th, 2012): Cape Breton Highlands and the Cabot Trail

by Jiyao Li

- 7:30 Wake up, breakfast
- 9:15 Pack up from Whycocomagh Camp Site and started on Cabot Trail
- 11:38 Arrive at Cape Smokey
- 12:54 Arrive at Ingonish Harbor
- 13:36 Arrive at abandoned gypsum quarries north of Ingonish
- 14:46 Lunch at Cabot Landing picnic area
- 14:00 Arrive at Stop 2: Arisaig Provincial Park
- 16:06 Visit Beluach Falls (Aspy Fault)
- 16:26 Aspy Fault Overlook
- 17:06 Anorthosite roadcut
- 17:29 Arrive at the west coast of Cape Breton Highlands
- 18:23 Arrive at Grande Falaise
- 20:00 Dinner at Coal Miner Cafe in Inverness, NS

We spent a whole day driving the Cabot Trail through the Cape Breton Highlands. Starting from the east coast and ending in the west coast, we observed different sequences of sedimentary rock and igneous rock, and visited the Aspy Fault which bisects the entire Cape Breton peninsula. We also enjoyed the fantastic view on the west coast and were lucky to see several large moose. Stop numbers corresponding to marked stops on the Nova Scotia Geological Highway Map (NSHM, Figure 2, next page) are indicated when available.

Stop 1: Cape Smokey (NSHM Stop #21): The rock type here is Later Ordovician granite. The lowlands along the coast are underlain by softer Carboniferous sedimentary rocks. On the north side there is a fault that juxtaposes reddish Carboniferous sedimentary rocks and the granite. The views of the east coast of Cape Breton are spectacular (Figure 1)!



Figure 1. Panoramic view looking east from Cape Smokey

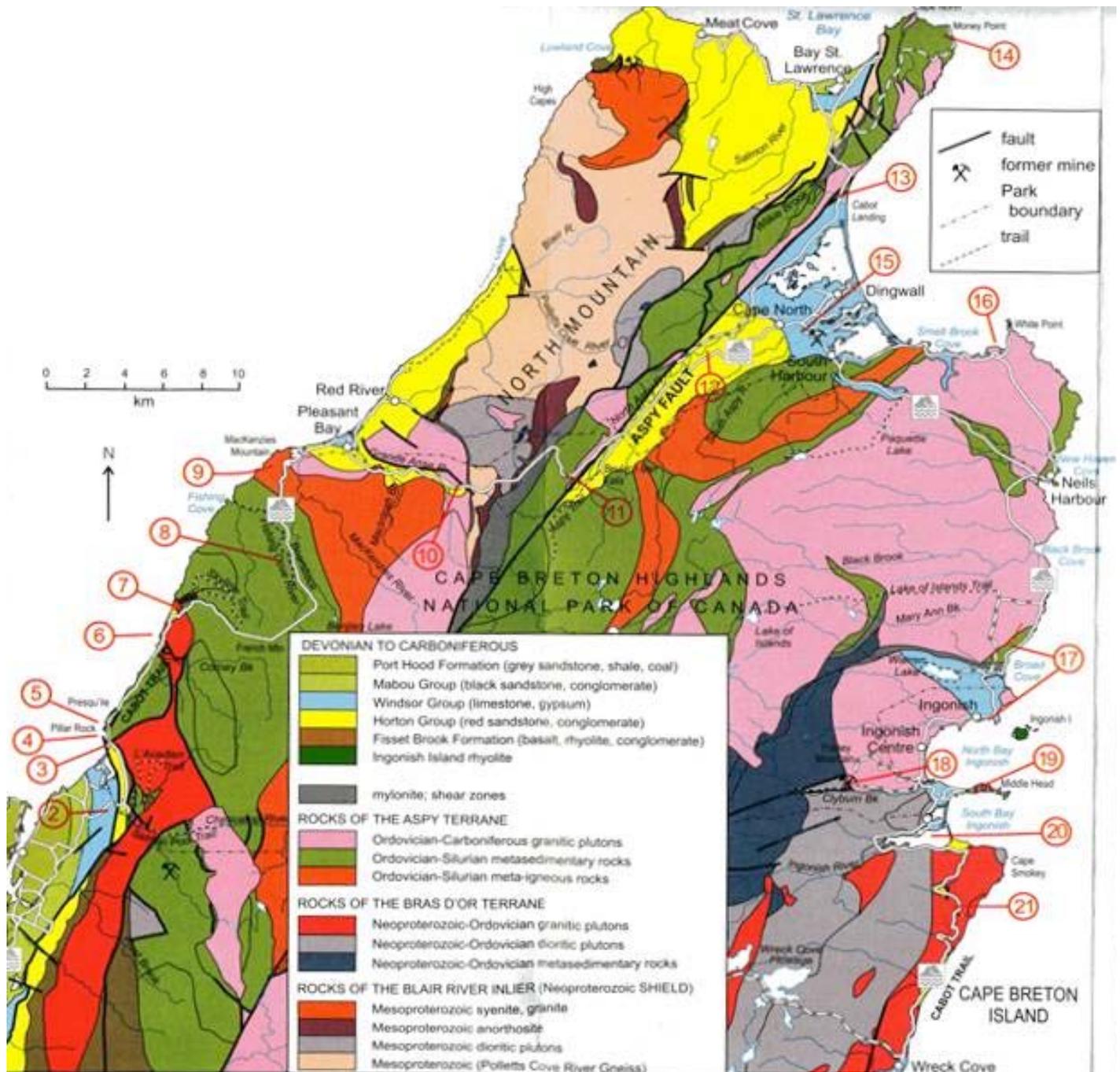


Figure 2. Geologic map of the northern part of Cape Breton Island.. The stop numbers in parenthesis are indicated on this map with red circles (From Nova Scotia Geological Highway Map)

Stop 2: Ingonish Harbor (NSHM Stop #17): There are two types of rock there. The sedimentary rock of the Windsor Group consists of limestone and gypsum. Above the Windsor Group is the Ordovician-Carboniferous granitic pluton. The sedimentary rocks form the harbor and the granitic rocks form the headlands (Figure 3).



Figure 3. Outcrop of the Windsor Group at Ingonish Harbor.

Stop 3: North Ingonish gypsum quarries (NHSM Stop #15): Windsor Group gypsum was mined at these locales from 1933 to 1954. Most gypsum in outcrop preserves their original white color but some sections are strongly weathered, showing yellow or even black color (Figure 4). A possible explanation could be mobilization of sulfate by groundwater or precipitation weathering, and reprecipitation of elemental sulfur, which carries a yellow hue. When a place is quite rich in sulfur, the color would be black.



Figure 4. Gypsum exposures at the abandoned mines. Left- yellow staining on gypsum outcrop, possibly from sulfur mobilization from weathering. Right- surface texture of a weathered gypsum fold.

Stop 4: Cabot Landing (NSHM Stop #13): While eating lunch, we observed an east-west oriented U-shape valley, formed by glacier motion. To the north we can see the steep escarpment of the Aspy Fault. This fault is thought to be the continuation of the Great Glenn Fault in Scotland, as both formed when Nova Scotia and Scotland were adjacent in the Devonian. The Aspy Fault has an offset of 15 m over the last 125,000 years (measured on a displaced wave cut terrace to the north of Cabot Landing), making it the most active neotectonic feature in Canada.

Stop 5: Beluach Ban Falls (NHSM Stop #12): The trail to the waterfall starts at Big Intervale warden station. This waterfall is on Aspy Fault and formed due to the difference in height due to differential erosion on either side of the fault. The waterfall comes down conglomerates of the early Carboniferous Horton Group (Figure 5).



Figure 5. Beluach Ban Falls along the Aspy Fault (left); Horton Group conglomerate (above).

In the plateau of the Cape Breton Highlands, moose are a common sight by the roadside. Drive slowly! Stopped cars along the side of the Highway are usually a sign of a moose sighting:



Stop 6: Aspy Fault (NSHM Stop #11): The Aspy Fault separates Carboniferous lowlands from older harder gneiss and granite of the highlands and forms a spectacular escarpment. There are also many hanging valleys that can be observed coming into the valley formed by the Aspy Fault. The Aspy river follows the trace of this structure. The views can be obscured if it is foggy. The highest pull-out is the best one! The road follows the north side of the fault; roadcut exposures of granites and gneisses are visible to the right (Figure 6).



Figure 6. Granitic exposures along the north flank of the Aspy Fault. The core of the Aspy Fault is located to the left of the picture, below the road.

We kept driving west and stopped to check out an outcrop of anorthosite, which is a rare type of intrusive igneous rock characterized by being 90-100% plagioclase (Figure 7). These rocks belong to the Blair River Inlier terrane.

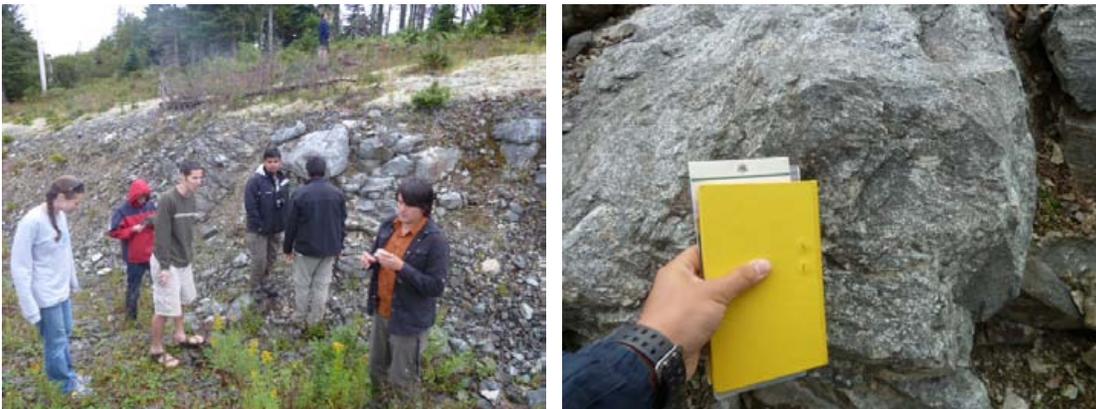


Figure 7. Anorthosite outcrop

Stop 7: Gulf of St. Lawrence (NSHM #9): At the top of the incline (Figure 8), we saw metamorphic and igneous rock like gneiss and pegmatite dikes. Horton group sedimentary units lie on the bottom and underlie Pleasant Bay. To the north we can see the headlands of High Capes formed by Neoproterozoic gneiss. On a nice day you can almost see the curvature of the Earth from this spot.

Figure 8. View of the Gulf of St. Lawrence from the west coast of Cape Breton



Stop 8: Grande Falaise (NSHM #3): Grande Falaise is a large cliff exposing Neoproterozoic pink granite on top of Devonian black volcanic rock (Figure 9). The older over younger relation indicates that this is a thrust fault. The fault is marked by a white line of sheared gypsum. The pink granite is cut by black dikes which were the feeder conduits to the volcanic rocks that were above the granite when they formed (but are now underneath it).

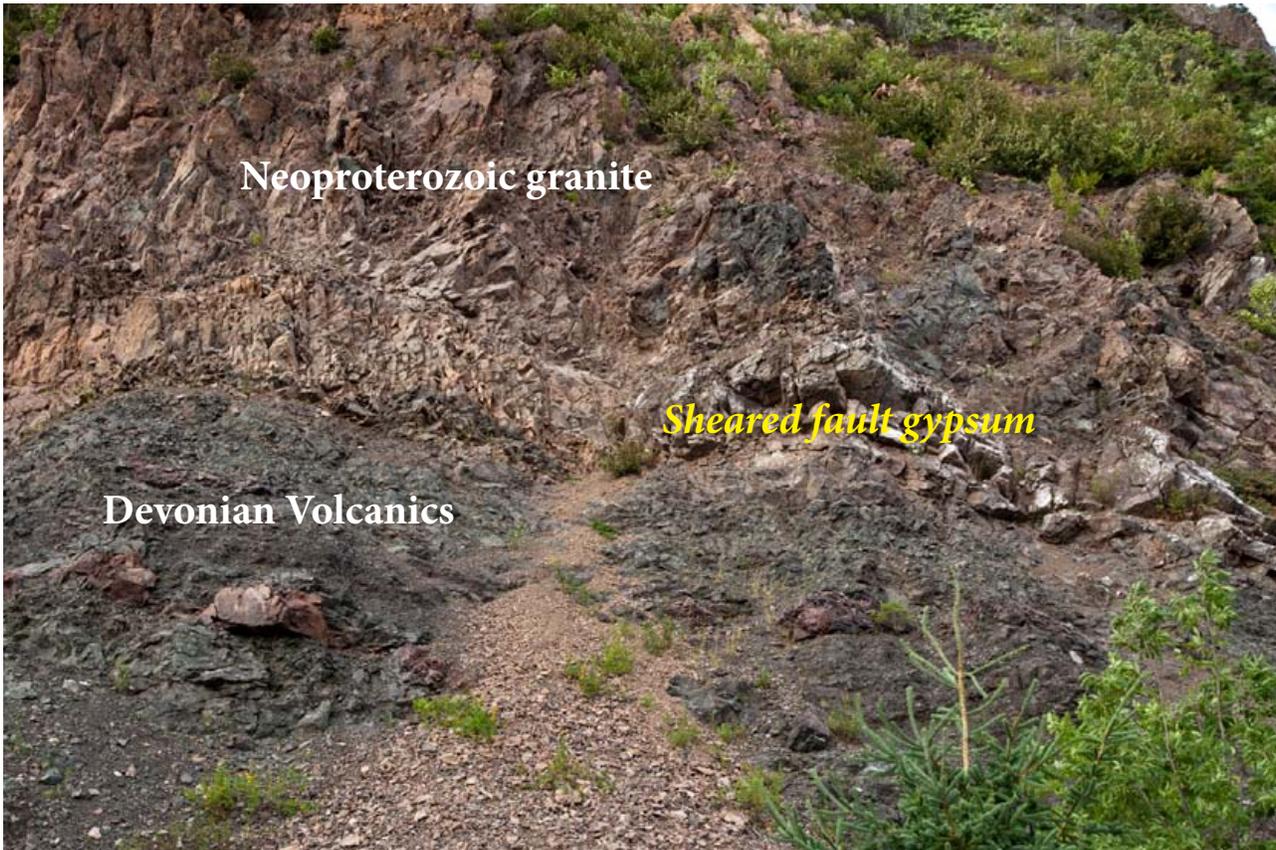


Figure 9. Large thrust fault exposure at Grande Falaise.

Day 8 (August 30th, 2012): Cape Breton Salt Diapirs & Bay of Fundy

by Chen Chen and Amelia Paukert

7:30 Wake up, breakfast, pack up tents

9:15 Leave Whycomomagh campsite, headed toward Mabou

9:40 Arrive at Mabou Harbor (mentioned in field guide, but not the correct location!)

10:05 Arrive at Finlay Point Harbor (correct location for diapirs!)

11:30 Leave Finlay Point Harbor, stopped at gift shop in Mabou

13:00 Lunch at Michael's Landing Park

14:00 Cross Canso Causeway, leaving Cape Breton

16:10 Arrive at tidal bore overlook (near Maitland)

16:40 Leave Maitland, headed to Burntcoat Head

18:10 Leave Burntcoat Head Park for campsite

17:29 Arrive at **Cape Blomidon Provincial Park** Campsite

Stop 1: Finlay Point Harbor & Coal Mine Point (46° 8'1" N, 61° 27'40" W)

Gypsum Cliffs are exposed on northern and southern sides of the harbor (Figure 1a). We parked the car in the harbor and walked to the northern part first, spending 40 minutes there.

The exposed formations at this outcrop were the Windsor Group (limestones, gypsums) and the Mabou Group (Figure 1b,c, Figure 2). The core was the Windsor Group. Salt started migrating up from the Windsor while the Mabou was depositing, so the Mabou had varying topography and accommodation during deposition. The salt migration caused deformation, and a few small normal faults formed to accommodate this deformation. The salt pushed up; the other rocks went down. There are mylonites present in the gypsum as a result of gypsum behaving more 'plastic-like' at lower temperatures than surrounding rocks. There are also coal seams present, but a dead whale carcass on the beach kept our time at the outcrop short.

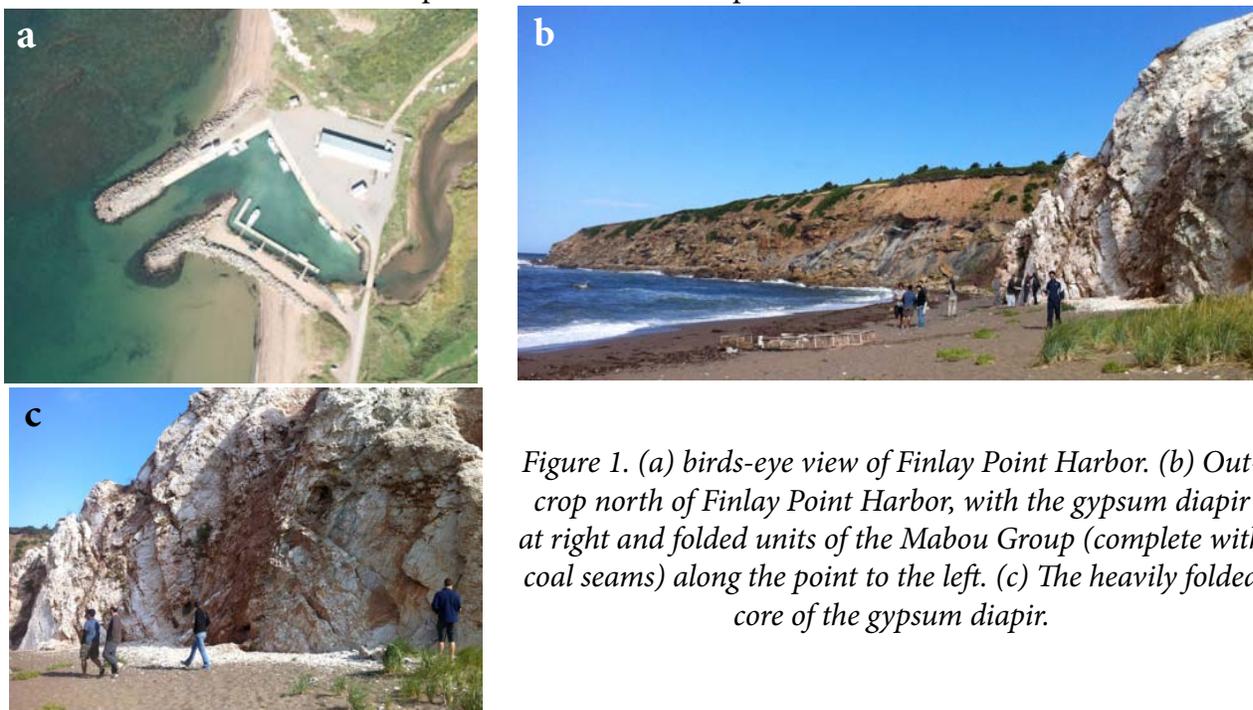


Figure 1. (a) birds-eye view of Finlay Point Harbor. (b) Outcrop north of Finlay Point Harbor, with the gypsum diapir at right and folded units of the Mabou Group (complete with coal seams) along the point to the left. (c) The heavily folded core of the gypsum diapir.



Figure 2. View of Finlay Point from Coal Mine Point. Windsor Group gypsum diapir core at right; tilted Mabou Group sediments along point. Coal seam is evident in dark layer in middle of the picture.

Next we drove to the southern side near Coal Mine Point (<5 min drive). We parked along the roadside next to house # 1081 on Mabou Mine Road (Figure 3a), then hiked down a small trail behind the house to the beach (the trail entrance is poorly marked, but is about 20 m behind the house). The hike took about 10 minutes. We spent 40 minutes on this side (Figure 3b). We saw recrystallized gypsum from salt from Huey, Duey, and Luey formations, and terrestrial plant fossils in fractured units of the overlying Mabou Group (Figure 4). Because the tide was high, access to Coal Mine Point was via a narrow gap between one of the gypsum cores and the waves; a good sense of wave timing or waterproof shoes are recommended to make it to the outcrops dry.



Figure 3. (a) the house at Mabou Mine Road with a trail to access Coal Mine Point. (b) view of Coal Mine Point from the beach to the north



Figure 4. Exposure of a terrestrial plant fossil in a loose boulder of the Mabou Group; pen for scale

Stop 2: Shubenacadie Canal(45° 15' N, 63° 27'7.2" W)

We parked the cars along the roadside and went down to the riverbank under the bridge. Unfortunately we missed the tidal bore because high tide already passed, but we observed water in the rivers rushing out to the bay. (Figure 5).



Figure 5. The Shubenacadie Canal. Already several meters of mudflats were exposed at right, even though it was not far past high tide.

Stop 3: Burntcoat Head Park (45° 18'46.8"N, 63° 48'18"W)

Burntcoat Head (Figure 6) is the site of the largest tidal range in the world. Along-shore outcrops contain Mid to Early-late Triassic rocks from tectonostratigraphic sequence 2. They are from the Evangeline member of the Wolfville formation. When they were deposited, this location was at approximately 10°N latitude. It was very humid during deposition, so the rocks have lots of bioturbation and bones. In this formation people found a fossil similar to a duck bill, molars, and claws like a shovel. There were no dinosaurs, but mammal-like reptiles were present. It was low tide when we got there, so we explored the outcrop for one hour. We found some fossil root features and evidence for tubeworms, but unfortunately found no major fossils despite an intensive search.

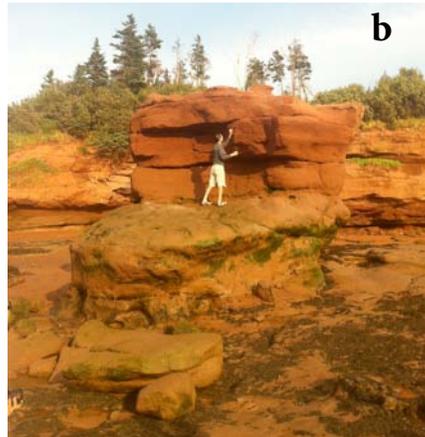


Figure 6. Burntcoat Head at low tide.

Day 9 (August 31st, 2012): Paddy's Island & Blomidon Park

by Marc Vankeuren

- 7:30 Wake up, breakfast
- 8:45 Leave Cape Blomidon Provincial Park
- 9:00 Observe tide at Delhaven Harbor
- 9:30 Arrive at Paddy's Island (end of N. Medford Rd)
- 11:15 Leave Paddy's Island
- 11:30 Visit outcrop north of Paddy's Island (across the bay)
- 12:20 Visit outcrop at entrance of Cape Blomidon Provincial Park (tide too high to walk up coast)
- 12:45 Lunch at the Lookoff, Gospel Woods Rd
- 14:30 Arrive at Pirate Pier (North Mountain Basalt + Scots Bay Member)
- 16:30 Depart Pirate Pier
- 16:45 Head to Port Williams for Supplies
- 17:30 Return to **Cape Blomidon Provincial Park** Campsite

Stop 1: Paddy's Island (45.157° N, 64.349 W)

Stop at Paddy Island to see lower units of Blomidon Fm. – tectonostratigraphic sequence #3; bottom of the Blomidon. At this locality, keep eyes open for dinosaur footprints. In outcrop, accelerated erosion here is due to isostatic rebound from latest glacial maximum retreat coupled with large variation in tides. Strata at this outcrop are north-dipping.

Rhythmic cycles on cliff outcrops are visible. Mud layers in outcrop show offset by normal faulting. 10-meter thick layers represent 100-kyr cycles, and thin layers represent 20-kyr cycles (Figure 1). Note, that several faults duplicate section in this area.



Figure 1. Lower units of the Blomidon Fm along cliffs at Paddy's Island stop.

Along the cliffs, strata layers here alternate from mud to sand to mud. Mud was deposited by standing water. Mud beds deposited over long distances from a lacustrine setting. Of note at the cliff outcrop is an example of a rupture of a forced fold. A normal fault cuts layers, with the strata south of the fault dipping up and into the fault, and strata on the north part of the fault dipping down into the fault – opposite of classic normal fault folding of layers (Figure 2a). Moving downsection, south, conglomerates begin to reappear with lake beds on top – coarser grained lake deposits. We recovered multiple crocodilian foot imprints from a sandstone bed from this unit (Figure 2b).

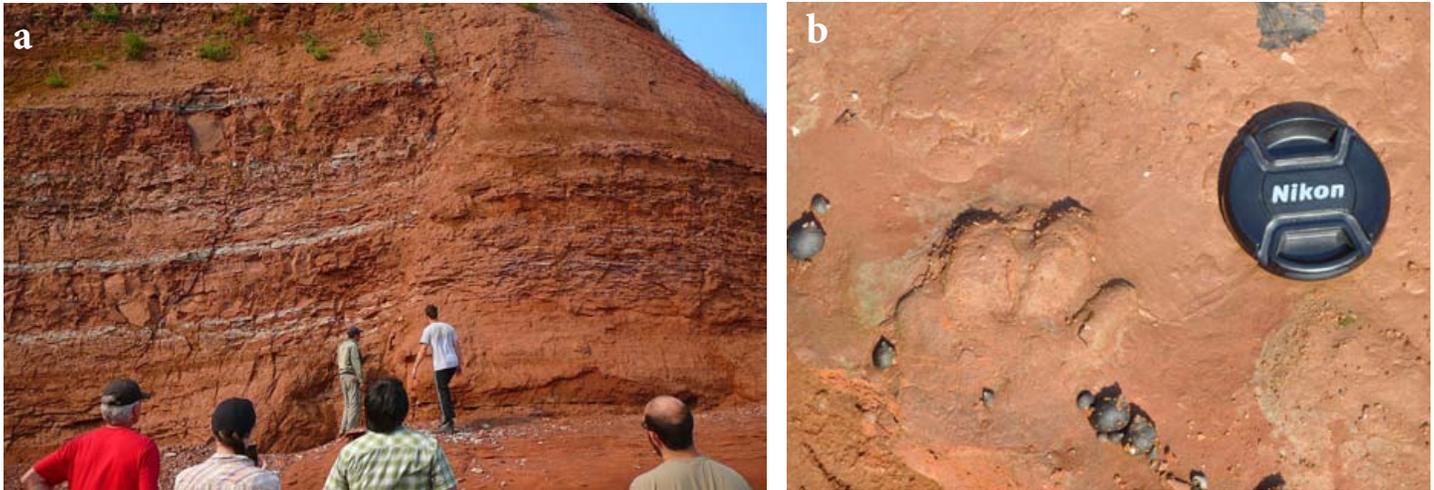


Figure 2. (a) Forced fold rupture, with strata to the left bending up into the fault zone, despite relative motion on the fault plane being downward on the left block. (b) Crocodilian track imprint in the Blomidon Fm. 2' x 3' sandstone block containing this and other tracks was retrieved from the Paddy's Island outcrop and returned to Lamont-Doherty Earth Observatory.

Stop 2: Beach ~2 km north of Paddy's Island

This outcrop location shows domino faults on the cliff face. Holes in the rocks here are the result of gypsum eroding out. Coarse sand is observed in fine sand on the outcrop – which depositionally make little sense – it forms as halite goes away during dissolution, creating a hummocky outcrop (Figure 3). Because of the dissolution, bodies of sand replace each other creating reverse stratigraphy. At this outcrop, shale breccia in sandstone is observed as the result of dissolution.



Figure 3. Lens of coarse-grained sandstone infilling between fine sand as a result of halite dissolution. Position of lens is denoted by black arrow.

Stop 3: Scot's Bay Member at Ross Creek (45.245°N 64.455°W)

Park along Ross Creek Road and walk out to the outcrop of the North Mountain Basalt along the coast (careful-very sharp and very slippery when wet. Do not go when it's raining, or when the tide is any higher than mid-tide). Green mud can be observed in the cracks of the basalt, the mud belongs to the Scot's Bay Member. Rocks in the cracks are jasper. There are three alcoves along the coast to the northeast (Broad Cove, Sheddley Cove and Phinney's Cove) with outcrops of the Scot's Bay Member, separated by terranes of the North Mountain Basalt (Figure 4a). Total time to transverse from the parking area to the third alcove and back is approximately two hours.

Broad Cove – Scot's Bay Member onlaps onto the basalt (Figure 4b). Bedding looks dish shaped because it filled the basalt depressions. Rocks are limestones. Stromatolites found on basalts. Centers of the stromatolites are filled with amethyst (Figure 4c).

Sheddley Cove – limestone on basalt. Used to be an excellent exposure but a storm event has created a berm covering most of the outcrop.

Phinney's Cove – Dinosaur footprints can be found at this locality in the sandstone. Limestones at this locality show knobby weathering, likely from stromatolites (Figure 4d).

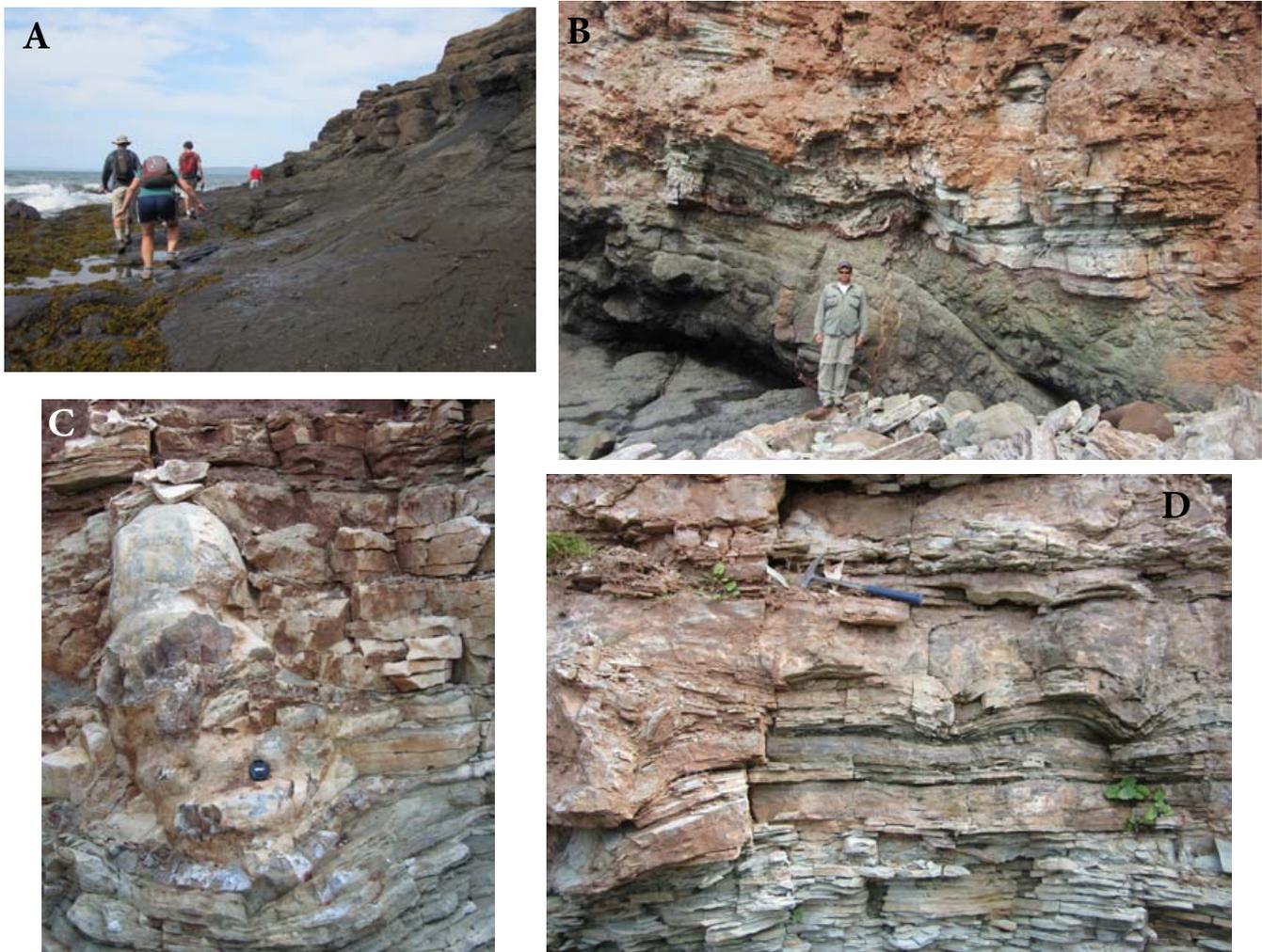


Figure 4. (a) North Mountain Basalt exposed along Scot's Bay; (b) Scot's Bay Member onlapping onto N. Mountain Basalt; (c) Stromatolite feature in Scot's Bay limestone; (d) knobby features in Scot's Bay limestone.

Days 10 and 11 (September 1st-2nd, 2012): Return to New York

by Jesse Farmer

7:30 Wake up, breakfast

9:00 Left Cape Blomidon Provincial Park

9:30 Arrive at Cape Split Park; hike for one last view of the Fundy Basin

11:30 Leave Cape Split Park and return to Cape Blomidon Provincial Park

12:00 Pack up campsite and Lunch

13:00 Depart Cape Blomidon Provincial Park for Digby, NS

15:30 Arrive at Digby, NS Ferry Terminal

16:00 Board Ferry for trip to St. John, NB

19:30 Arrive in St. John, NB

20:30 Stop and get dinner along Rt. 1 in Canada

21:15 Arrive at US-Canada Border

22:15 Arrive at **Boidi Motel**, Boidi, ME

6:30 Wake up, breakfast

7:30 Depart for New York

16:45 Arrive at Lamont-Doherty Earth Observatory

17:30 Return Rental Cars in New York City



VI. Campsite and Lodging Information

Lamoine State Park

23 State Park Road, Ellsworth, ME, USA

(207) 667-4778

Reservation recommended (<http://www.campwithme.com/doc/parks/reservations/app/fees.htm>)

Group Site: \$25 reservation fee and full rate due on arrival

Outhouses and potable water at campsite; central location with showers and toilets

Seaside Campground

234 Main St., St. Martins, New Brunswick, Canada

(506) 833-4413

\$27 per site, four sites for the group

Wireless internet across campgrounds, pool, showers and toilets

Five Islands Provincial Park*

Bentley Branch Rd, Hwy 2, Five Islands, Nova Scotia

(902) 254-2980

Large group sites, potable water and outhouses near site. Central location with showers and toilets.

Whycocomagh Provincial Park*

97 Provincial Park Rd, Whycocomagh, Nova Scotia

(902) 756-2448

Potable water at campsite. Group site is located on a hillside and has significant slope. Central location with showers, toilets, and laundry

Cape Blomidon Provincial Park*

3138 Pereau Rd, Canning, Nova Scotia

(506) 833-4413

Potable water and outhouses at campsite; central showers and toilets. Group sites are immediately to the left after entering the campground

Boidi Motel

12 Houlton Rd, Baileyville ME 04694 USA

(207) 454-7403

Four rooms for one night: \$280

Located next door to a full service gas station with diner

*Provincial Park Group Site Fees:

\$18.36 per day without flush toilets/showers

\$24.48 per day with flush toilets/showers

+\$2.17 per person for groups larger than 6 people

VII. Trip Budget

EXPENSES (Actual)	Car 1	Car 2	Car 3	Car 4	Total
Lodging - campsites	418.03			126.98	\$ 545.01
Entrance Fees	364.6				\$ 364.60
Food	1923.03				\$ 1,923.03
Gas and tolls	598.56	498.58	548.31		\$ 1,645.45
Supplies	182.36				\$ 182.36
Transportation	672.48				\$ 672.48
Credit card/transaction fees	98.60	8.42		1525.97	\$ 1,632.99
Car Rental	1047.70	1251.38	1007.13		\$ 3,306.21
Total Expenses	5305.36	1758.38	1555.44	1652.95	\$ 10,272.13

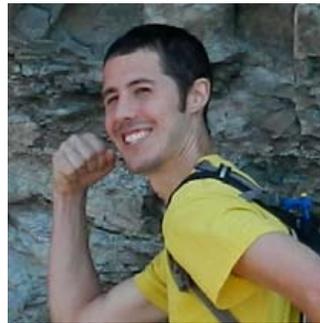
Fees collected from 12 students

\$ (3,600.00)

Net to DEES

\$ 6,672.13

VIII. Participant Gallery



STUDENTS: (left to right, top row to bottom)
Rafael Almeida, Chen Chen, Jacob Eichenbaum-Pikser, Jesse Farmer, Anna Foster, Alison Hatman, Jiyao Li, Alexander Lloyd, Pritraj Moulik, Amelia Paukert, Marc Vankeuren, Mike Wolovick



Professor Paul Olsen

IX. Acknowledgements

The success of any field expedition is a reflection of contributions of the individuals who donated their time and assistance throughout the planning and execution stages. The participants of the 2012 Arthur D. Storke Memorial Expedition to Nova Scotia would like to thank:

- The Storke Endowment Field Fund, for making this trip possible!
- Professor Peter deMenocal, chair of the Department of Earth and Environmental Sciences, for his continued support of graduate student field expeditions;
- Amelia Paukert, for getting the ball rolling on planning a 2012 Field Expedition;
- Professor Mike Melchin of St. Francis Xavier University, who graciously took a day of his time to serve as our field leader during our visit to the Arisaig Group on Day 6. His in-depth knowledge of the stops and the units in outcrop were much appreciated by all.
- Rafael Almeida (cowgirl below) and Anna Foster (cowboy below), who took the reins on organizing the logistics of the field trip, getting departmental approval, organizing the Summer Seminar series (so we all knew what we were looking at in the field), and pretty much organized every single detail that required organization at any point before, during, or after the trip. During his ~~fifteen~~ five years at Lamont, Rafael has been a major proponent of the graduate student field trips and a tireless organizer of multiple expeditions, and has left quite the legacy of successful, educational, and downright fun trips in his wake.
- And finally, to our fearless leader, Paul Olsen (cactus below). It was both a privilege and an honor to join Paul in this expedition to the backyard of much of his life's work. Intellectually, you could not have picked a better guide; Paul knew every stop and outcrop like the back of his hand, and was more than willing to answer the litany of questions that we fired away. In the field, Paul earned our respect from the first day forward, taking us on beautiful six-mile hikes with a half a bottle of diet Mountain Dew, making it first to each stop and barely breaking a sweat by the end. And socially, Paul greatly enhanced the fun of the trip. Whether it was our late-night chats by the campfire, or stopping to take in some of the beautiful views in Nova Scotia, or going off-script for a photo shoot (like below), Paul maintained a relaxed, collegiate attitude for the whole trip, enhancing the camaraderie of the group. Suffice to say, all of us would be honored to join Paul on another field expedition.

