Central Atlantic Conjugate Margins Conference | Halifax 2008

Field Trip Guidebook

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

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CENTRAL ATLANTIC CONJUGATE MARGINS CONFERENCE

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Tropical to Subtropical Syntectonic Sedimentation in the Permian to Jurassic Fundy Rift Basin, Atlantic Canada, in Relation to the Moroccan Conjugate Margin

Saturday August 9th –
Tuesday 12th 2008

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PREFACE

This field trip examines the spectacular outcrops of Late Permian, Triassic, and Jurassic age Fundy rift basin strata exposed along the shores of the Bay of Fundy, in New Brunswick and Nova Scotia, Canada. We will meet the evening of August 8, 2008 at 1 Market Square, St. John, New Brunswick, Canada and begin the fieldtrip proper on the morning of August 9, 2008 at the same location. We will travel NE through eastern New Brunswick on August 9th examining field stops, ending that day in Parsborro, Nova Scotia. We will then examine stops along the North shore of the Minas Basin segment of the Bay of Fundy on August 10th ending again in Parsborro. August 11th and 12th will be spent traveling to the south shore of the Minas Basin where we will examine field stops staying the night of August 11 in Wolfville and end the field trip on August 12th in Halifax, Nova Scotia.

This field guide is divided into two parts. Part 1 of this presents a general description of early Mesozoic rift basins in Eastern North America and Morocco, providing the context for the field stops. Part 2 is a description of the field stops themselves with comparisons to contemporaneous basins in Morocco.

SAFTEY

For personal and group safety we ask all participants to read and heed the following safety related procedures. We ask for your cooperation and common sense in making this a safe and enjoyable field trip for everyone. Thank you.

1. ROCK HAMMERS: Some sites we are visiting (Wasson Bluff, Five Islands) are protected and it is illegal to use rock hammers without a Nova Scotia Heritage Permit. Of course, in general please use caution when hammering: be aware of people around you, use controlled downward blows, and do not hammer indiscriminately.

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

3. HARD HATS: Hard hats are recommended anywhere you intend to look at rocks where there are cliff faces or overhangs. We will have a supply of hard hats for use by field trip participants.

4. CLIFFS AND FALLING ROCKS: We will be visiting sites that experience tremendously high rates of erosion, which result in the possibility that exposed rock faces will be unstable, therefore posing a major hazard on field trips. Avoid unstable waste rock piles or overhanging cliffs and watch for people below you on slopes.

5. TIDES: The Bay of Fundy has some of the highest tides in the world, with a vertical range of 15 meters, which is great for exposing fresh outcrop. However, always note whether the tide is coming in, or going out, and ensure that access routes will not be cut off with an incoming tide. Inter-tidal rock exposures can be very slippery, please be very cautious when walking on wet outcrop or cobbles, and especially on algae covered surfaces.

6. FIRST AID / MEDICAL CONDITIONS: Several First Aid kits will be located in the bus/vans and with a leader on site. Participants with valid First Aid certificates are encouraged to identify themselves at the beginning of the field trip. Field trip participants with medical conditions (allergies, diabetes, etc.) may wish to advise the field trip leaders prior to departure.

7. IN THE UNLIKELY EVENT OF AN EMERGENCY: Call 911. Some areas in Nova Scotia have poor cellular phone coverage, so it may be necessary to use a pay/private phone, and therefore we have a sattelite phone to be used exclusively in the case of an emergency.

PART 1: Context of the Rift Basins of the Fundy and Moroccan Conjugate Margins

Introduction

One of the most characteristic features of the rift basins of the central Atlantic margin (Fig. 1) is thick sequences of red mudstone of Late Triassic (and sometimes Early Jurassic) age. In the more northern basins of present day Maritime Canada, Morocco, Iberia, Central Europe and Great Britain, these red mudstones are often associated with evaporites, which sometimes reach kilometers in thickness. Although these evaporites are often of critical importance to the petroleum system, their overall stratigraphic, climatic, and temporal context as well as that of the associated clastics are surprisingly poorly understood.

During this field trip we will examine first hand outcrops of the largest of the exposed North American contingent of these rifts the Fundy basin and make direct comparisons with its sister basins on the conjugate margin of Morocco (Fig. 2), as well as basins, further south in North America. In particular, the Fundy basin most strongly resembles the Argana and associated Essaouria basins in western Morocco and its contiguous shelf, but there are also profound similarities to the High Atlas basin complex (HABC - High Atlas and Central High Atlas basins), and these will, therefore, be the principle focus of this field guide. However, we will also discuss the trend to the north and east to larger marine involvement in the Moroccan basins.

These outcrops and comparisons will be examined with the goal of understanding the development of the Petroleum system with these basins, including the synrift development (or not) of source rocks, reservoirs, seals, and tectonic history, as well as the possible role of pre- and postrift source and seals.

Synrift Tectonic Sequences

All of the synrft conjugate margin basins have sequences that show strong vertical changes in overall facies reflecting an interplay of large-scale tectonic and climatic changes with time. The basic outline of these has long been clear (Van Houten, 1977; Manspeizer, 1988). In specific, the basin sections are made up of 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 et al., 1989) (Figs. 3, 6). Similar asymmetrical vertical sequences have been observed in many other continental rifts (Lambiase, 1990). Such sequences have been explained as the result of a specific history of tectonic events (Manspeizer & deBoer, 1989), as a consequence of sequential filling of hydrographically linked basins (Lambiase, 1990; Contreras & Scholz, 2001), and as the necessary result of the filling of a widening half graben bounded by a normal fault (Schlische & Olsen, 1990, Schlische, 1993; Contreras et al., 1997; Contreras & Scholz, 2001).

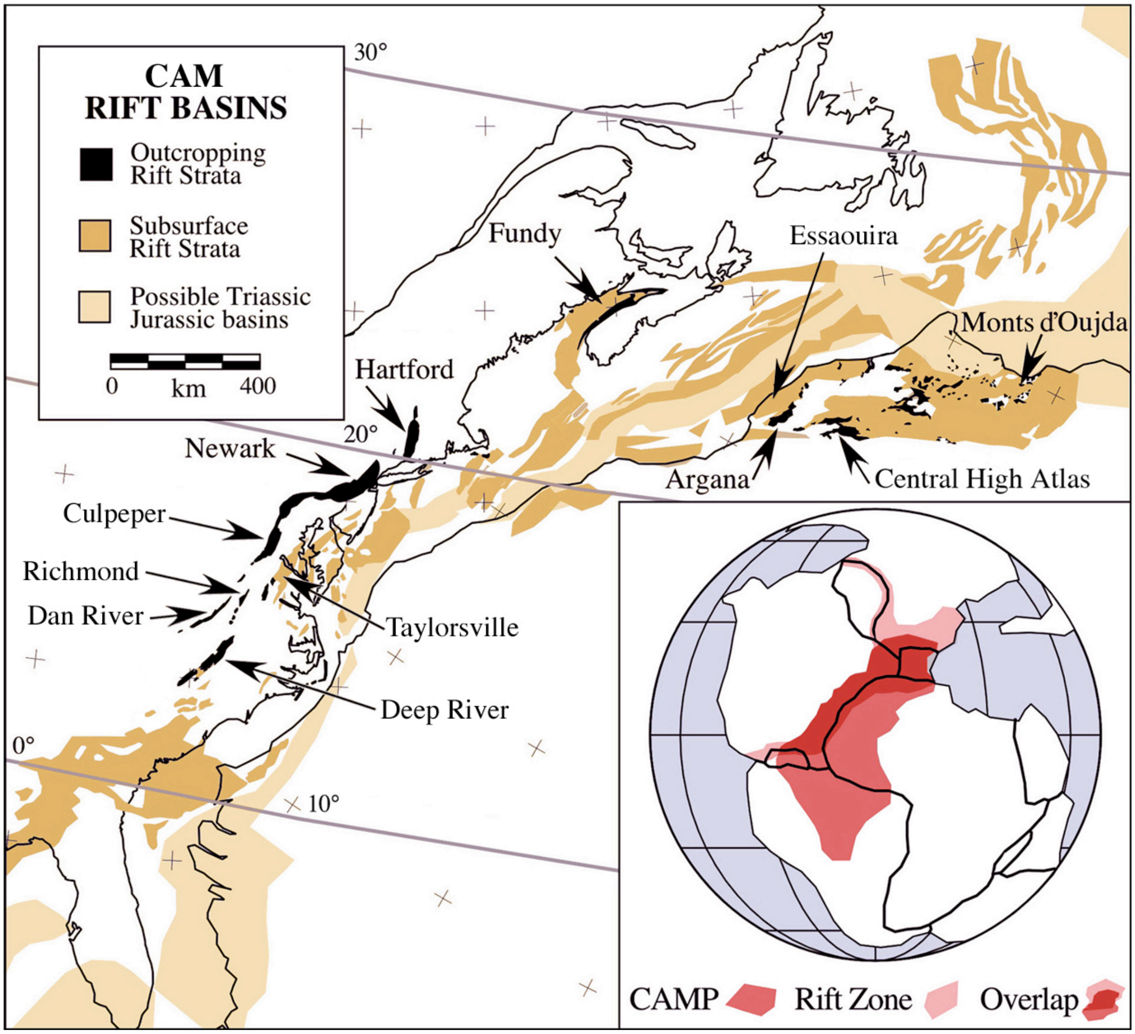


Figure 1: Central Atlantic Margin rift basins. Adapted from Olsen (1997).

Schlische & Olsen (1990) proposed a simple quantitative model in which the rate of subsidence, the rate of inflow of sediment into the basin, and the basic geometry of the basement components of the basin remain constant. As a simple consequence of the increase of the area of the depositional surface though time, the full tripartite lacustrine sequence is produced. This basin-filling model predicts the sediment accumulation rate history of the basin; once lacustrine conditions are established, accumulation rate should decline exponentially.

The first quantitative data on long-term sedimentation rates comes from the NBCP and ACE cores (Olsen et al., 1996a,b). Comparison of the sedimentation rates of the Triassic-age part of the basin section with the predictions of the Schlische & Olsen (1990) model shows some first-order differences (Fig. 3). Specifically, a simple exponential decline in sedimentation rates is not seen; however, the facies transitions predicted in the Schlische & Olsen model are confirmed. A much more realistic, although more complicated, model of evolving rift basin geometry is presented by Contreras et al (1997) (Fig. 3). This is a three dimensional finite-difference model based on the fault-growth models of Cowie & Scholz (1992). Contreras et al (1997) show that for a half graben bounded by a single normal fault the accumulation-rate curve predicted by the Schlische & Olsen (1990) model occurs only at the depocenter of the basin, adjacent to the border fault. All other parts of the basin have different accumulation-rate curves.

An accumulation-rate curve similar to that seen in the Newark basin data (Fig. 3) is obtained in the model if the basin is sampled by laterally offset drill-hole data as were obtained in the coring project.

Viewed in light of the above models, each tripartite sequence probably represents a different tectonic episode, basically a rejuvenation of extension or an acceleration of extension. This hypothesis can be expressed as an extension of the Schlische & Olsen (1990) or Contreras et al (1997) models. Thus, relatively low rates of extension and basin subsidence allow the available sediments to fill the basin to the outlet, resulting in a hydraulically open basin with through-flowing streams and a coarse-grained basin fill. Slow rates of extension should result

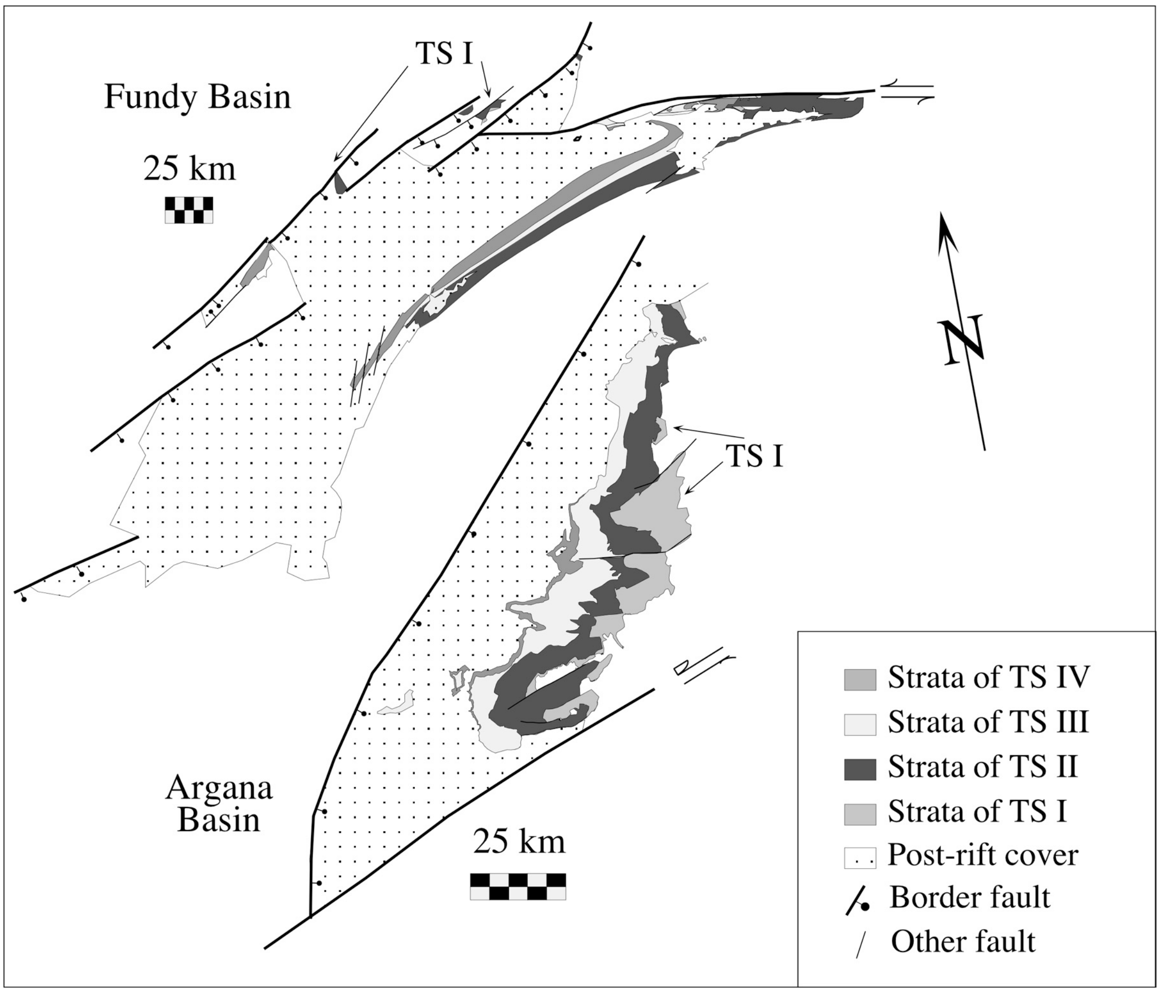


Figure 2: Fundy and Argana basins showing distribution of tectonostratigraphic sequences. From Olsen, 1997.

in similarly slow rates of footwall uplift. Hence, it could have been relatively easy for inflowing or outflowing rivers to breach the footwall. The fluvial portions of the Stockton Formation of the Newark basin (Smoot, 1991) may be an example of streams leaving through the footwall, whereas the New Haven Formation of the Hartford basin (Hubert et al., 1978) may be an example of a fluvial system entering the basin though the footwall. When extension rates increased, basin floor subsidence and tilting increased. As a consequence, the basin expanded at a rate faster than it could fill, and a closed drainage basin resulted. When sufficient water became available, a lake could form. Smoot (1991, 2008), on the other hand, feels that an autogenic reduction of fluvial gradient due to clogging of the outlet resulted in the fluvial-lacustrine transitions. As long as extension rates remained high, the distance between the outlet and the basin floor would continue to increase as the area of the depositional surface expanded; and as long as there was sufficient precipitation, lacustrine conditions would be maintained (Schlische & Olsen, 1990). High extension rates would lead to high rates of footwall uplift, and the basin would be sourced from either the hanging wall [upper Stockton and Lockatong formations, Newark basin (Glaeser, 1966) or the axis (Passaic Formation, Newark basin; Smoot, 1991).

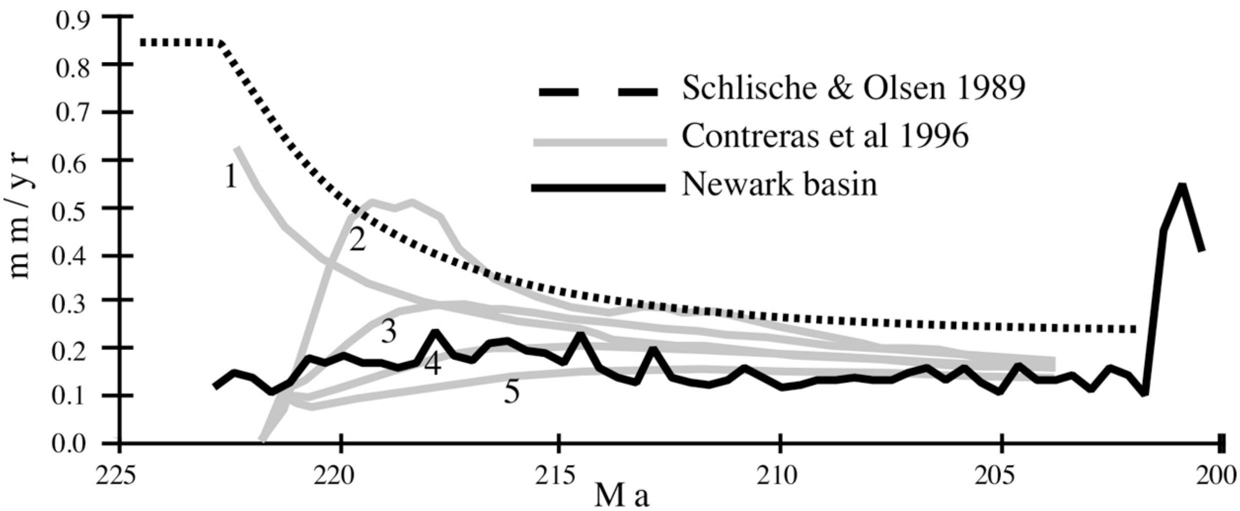


Figure 3: Comparison of accumulation rate predictions of models of rift basin evolution with data from the NBCP cores. Numbers 1-5 refer to predictions at increasing distances from the border fault. From Olsen (1997).

As extension decreased, the basin would tend to fill, eventually reestablishing fluvial conditions. Footwall uplift would correspondingly decrease and the footwall could again be breached by streams, perhaps capturing large drainage areas on the back of the footwall uplifts (upper Portland Formation, Hartford basin). If subsidence stayed low or stopped, the outlet would be cut into, and eventually the basin fill would begin to erode. Thus, the full cycle of an extension pulse would tend to produce one of the tripartite lacustrine sequences. If the magnitude of the extension pulse is relatively small, then a short-lived excursion though a tripartite sequence might be produced. If the pulse of accelerated extension followed a period of slow extension or erosion or if it was of large enough magnitude to produce erosion of footwall sediments before they could be onlapped by the new tripartite sequence, then an unconformity might develop that would separate the two sequences along the up-dip portions of the basin. However, as long as the basin remained hydraulically closed, a conformable relationship between two successive sequences would still be expected at the depocenter of the basin. In addition, an abbreviated yet strong pulse might only produce a fluvial sequence. Multiple tripartite sequences and unconformity-bound fluvial sequences are observed in many Atlantic conjugate margin basins, which suggests that their basin fill can be organized into discrete tectonostratigraphic sequences.

Tectonostratigraphic Sequences of the Fundy and Moroccan Basins

Traditionally, the main organizing principle within basin sequences has been the recognition of mappable lithostratigraphic formations defined on one or a few criteria, such as color or grain size. However, industry and scientific drilling and seismic profiling, as well as reexamination of outcrops, has shown that four basic tectonostratigraphic sequences (TS) can be recognized that extend over most, if not all, of the early Mesozoic rift basins (Fig. 4). These are sequences in the sequence stratigraphic sense (see Christie-Blick & Driscoll, 1995); they are mostly but not always unconformity bound and probably resulted from significant changes in the rate of extension (rather than sea level changes, which in these basins is nearly irrelevant). Age is the criterion used here to group sequences in isolated basins into individual categories. These categories are as follows: TS I, initial synrift sedimentary rocks of Late Permain age; TS II, early synrift sedimentary rocks of possibly Middle Triassic to early Late Triassic age (Anisian to Carnian ages); TS III, middle synrift sedimentary rocks of Late Triassic age (Norian and Rhaetian); and TS IV, late synrift sedimentary and volcanic rocks of Early Jurassic age (Hettangian Age), plus the overlying late synrift sedimentary rocks of Hettangian to possibly Pliensbachian (Early Jurassic) age (Fig. 2). These tectonostratigraphic sequences provide the framework for understanding the large-scale tectonic events that influenced basin evolution as well as a logical structure for the discussion of the rift sequences themselves.

The Fundy Basin of the Maritime provinces of Canada (Fig. 2) is a very large half graben complex, with its main bounding faults on the northwest and north. The exposed portion of the Argana basin of Morocco is much smaller basin than the Fundy basin and it is a southeastern extension of the large Essaouria basin (Fig. 2), the bounding faults of which are not as well constrained. The HABC (Fig. 1) is closely comparable to the Argana basin in stratigraphy and may in fact have been connected to it, prior to the uplift associated with the Alpine orogeny.

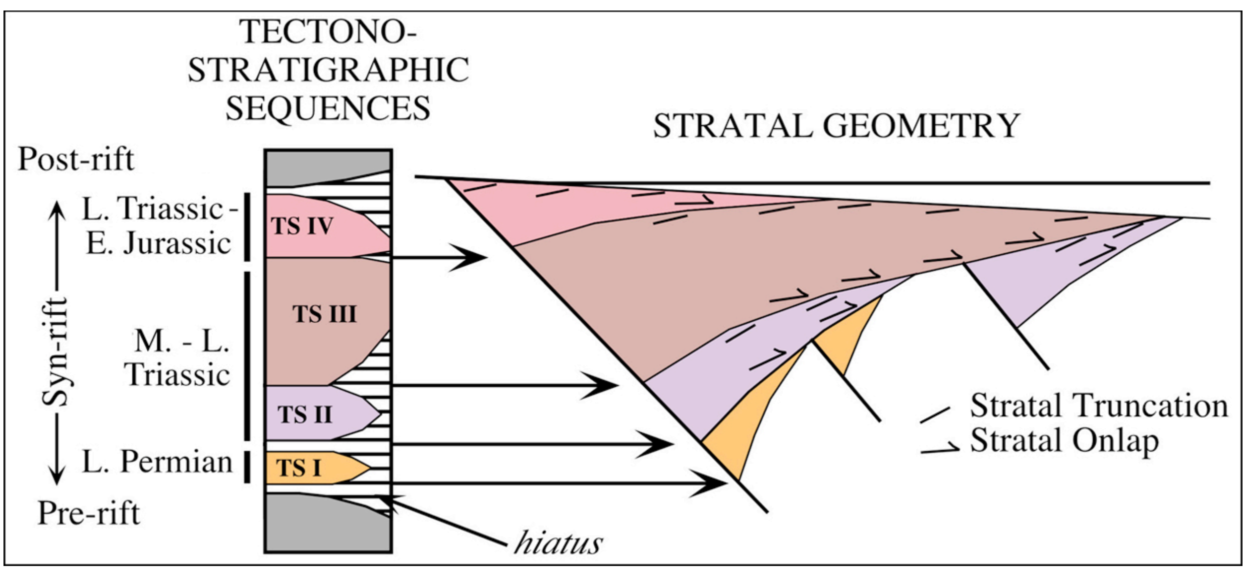


Figure 4: Tectonstratigraphic sequences of rift sequences in the Fundy and Moroccan basins (Modified from Olsen, 1997).

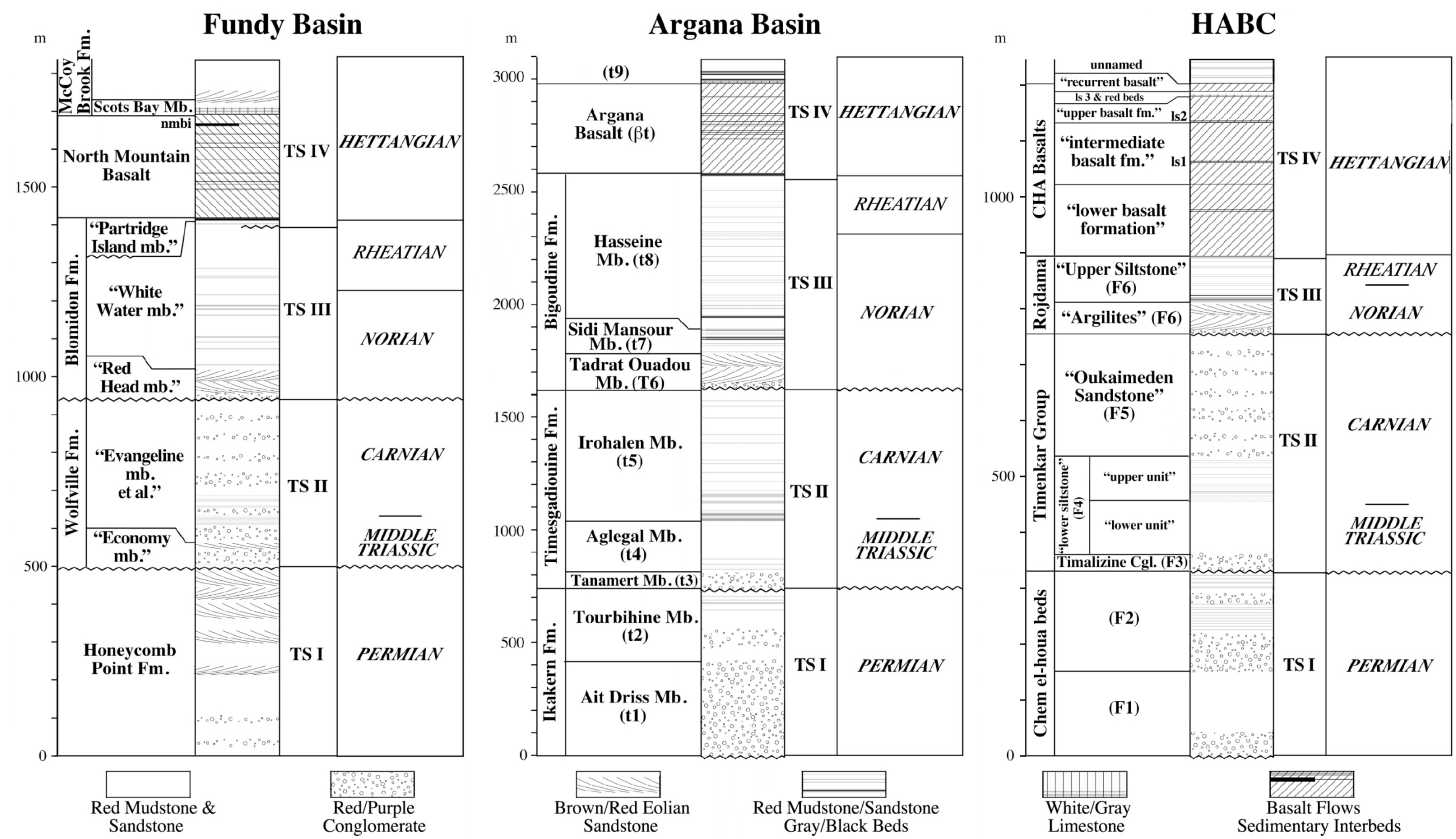


Figure 5: Tectonstratigraphy, lithostratigraphic divisions, and age of the Fundy, Argana, and HABC basins (modified from Olsen et al.,2003; El-Arabi et al., 2007; Benaouiss et al., 1996; Whiteside et al., 2008). Note that the thicknesses of all units vary dramatically across each of the basins.

The Fundy, Argana, and HABC basins have remarkably similar stratigraphic sections that are divided into four tectonstratigraphic sequences that are at least locally separated by synrift unconformities, termed, from the bottom up, TS I – TS IV (Olsen, 1997; Olsen et al., 2003a) (Figs. 3, 5). Most of these tectonostratigraphic sequences have been identified in the other rift basins of the North American and African continental margins.

Tectonostratigraphic Sequence I

TS I comprises initial synrift sedimentary rocks that consist of strongly wedge-shaped sequences in small basins with beds fanning markedly towards faults as seen in the Argana basin) (Fig. 3). TS I has been recognized so far only in the Fundy basin and the Argana and associated basins of Morocco. The sequence rests with a profound unconformity on underlying prerift rocks and is separated from younger synrift strata by an angular unconformity (Stops 1.1, 1.2). TS I consists mostly of fluvial and aeolian rocks deposited under an overall arid environment as seen in isolated fault blocks along the border fault of the Fundy basin in New Brunswick Canada (Stops 1.1-1.3). Outcrops consist of alternating mostly fluvial and aeolian dune sequences and minor red lacustrine mudstone (Hubert & Mertz, 1984; Nadon & Middleton, 1985) (Stop 1.1). Based on paleomagnetic data these comprise the oldest dated strata in the Newark Supergroup. Spectacular unconformities separate TS I and TS II of Morocco (Fig. 6), and these strata contain tetrapods of undoubted Late Permian age.

The Honeycomb Point (and probably the Lepreau Formation) of New Brunswick and the Ikakern formations (Stops 1.1 and 1.2) comprise TS I in the Fundy and Argana basins respectively (Fig. 4) (Olsen, 1997). TS I is probably represented by the Cham-El-Houa beds (El El Arabi et al., 2006) (F1 and F2 of Biron, 1982 and Biron & Courtinat, 1982) of the High Atlas basin complex (Fig. 5), although age-diagnostic data are lacking. The presence of undoubted examples of the hammer-headed amphibian *Diplocaulus* and a pareiasaur from the Ikakern (Jalil, 1996), as well as preliminary paleomagnetic data (Kent et al., in prep) from the Honeycomb Point and Ikakern formations demonstrate a Late Permian age. The Honeycomb Point Formation is distinctive in having well developed eolian strata, while the Ikakern Formation and Cham El Houa beds do not, owing probably to the more southerly and tropical location of these Moroccan units in Late Permian geographic coordinates (Olsen et al., 2000), with Pangea being rotated clockwise relative to its orientation in the Late Triassic (Kent and Muttoni, 2003) (Fig. 7).

While there is evidence of deposition of the Ikakern Formation and the Cham El Houa beds in an active extensional basin setting (Medina, 1991; Fig. 5), such evidence for the Honeycomb Point Formation is not so clear. In fact, the Honeycomb Point is almost lithologically identical to the apparently coeval Cap aux Meules Formation (Brisbois, 1981; Tanczyk, 1988) of the presumably non-rift Magdalen basin the Gulf of St. Lawrence, which seems to have been deposited as a very broad sheet. The Honeycomb Point Formation could conceivably just be an outlier of this sheet, fortuitously preserved within the Fundy Rift. However, the known thickness of the Honeycomb Point (~1 km) Formation and especially its presumed equivalent the Lepreau Formation (~2.7 km) is vastly greater than known thickness of the Cap aux Meules Formation suggesting a different mode of accommodation space, such as rifting. To our knowledge TS I has never been penetrated in the subsurface in the Fundy basin and its distribution is unknown in the more southern Newark Supergroup basins of the United States.

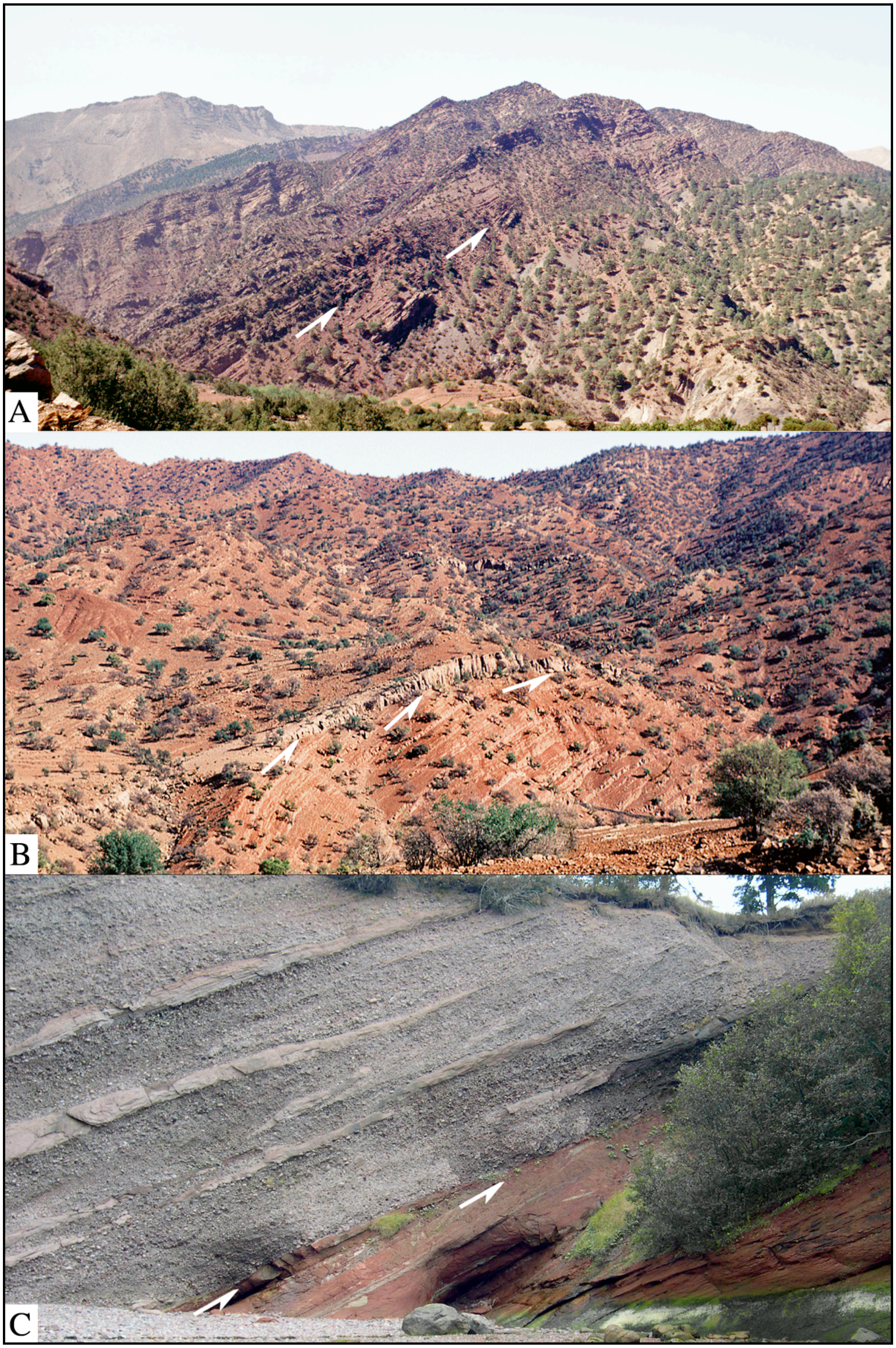


Figure 6: Unconformities between TS I and TS II: A, Aït Tamlil, High Atlas basin, Morocco; B, Argna Basin, Morocco; C, Fundy Basin, St. Martins, New Brunswick, Canada (Stop 1.2).

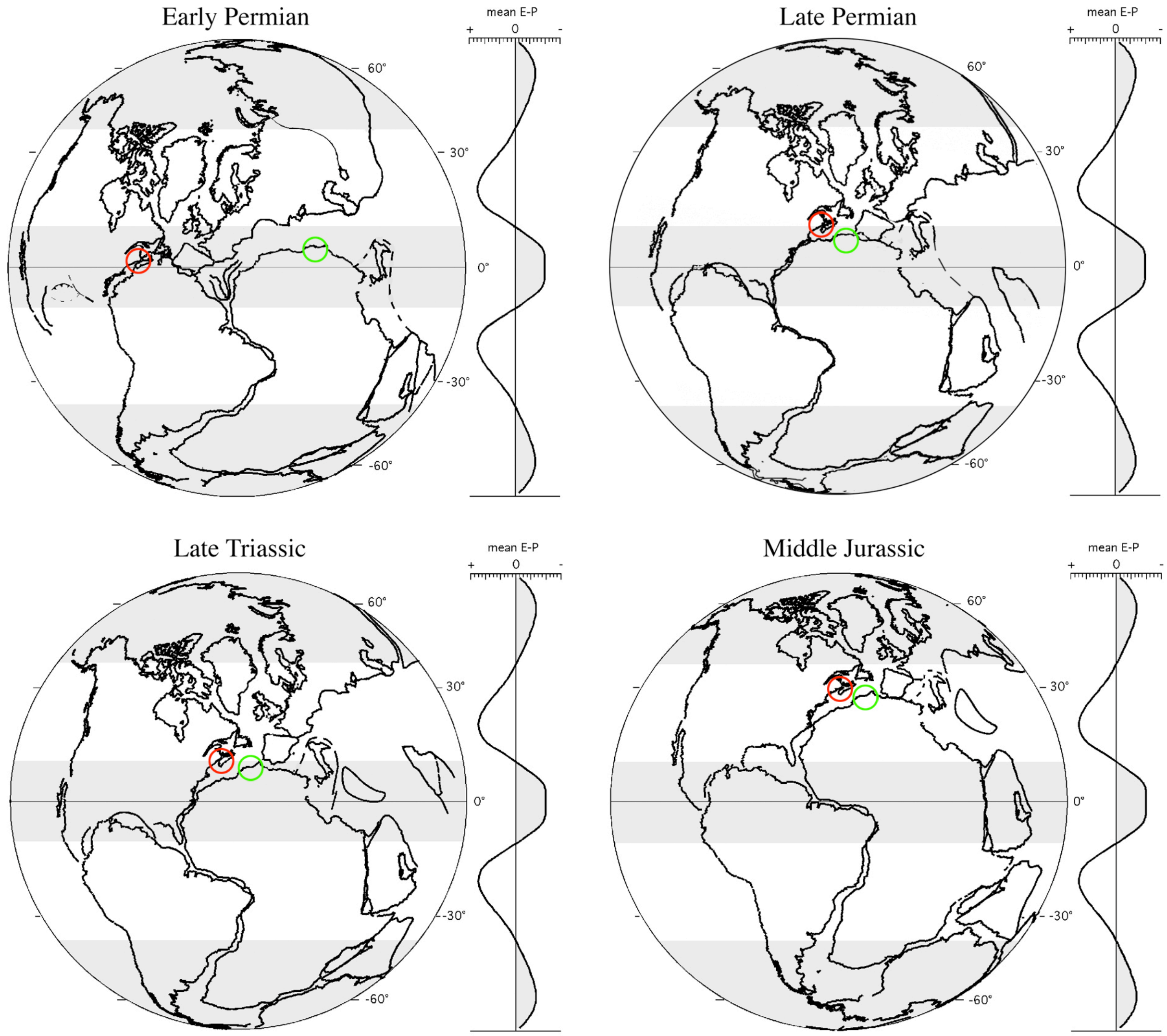


Figure 7: Plate position reconstructions for the Early Permian through Middle Jurassic showing gradual counterclockwise rotation of central Pangea. Shown for reference is the mean zonal variation in evaporation minus precipitation (E – P) for the modern land plus ocean surface (Crowley & North 1991), which has been further averaged over the Southern and Northern Hemispheres. The locations of the Fundy and Moroccan basins are indicated by red and green circles, respectively. Tethyan tectonics are highly diagrammatic. Early Permian (~275 Ma) reconstruction resembles Pangea B (modified from Kent & Muttoni, 2003). Late Permian-Early Triassic reconstruction (~252 Ma) resembles Pangea A-2 of Van der Voo and French (1974) (modified from Muttoni et al. (1996). Late Triassic (~225 Ma, Early Norian) reconstruction also resembles Pangea A-2 but with the North American-African suture rotated more counterclockwise (from Kent and Muttoni, 2003). Middle Jurassic (~175 Ma) reconstruction that resembles Pangea A-1 of Van der Voo and French (from Kent and Muttoni, 2003).

Tectonostratigraphic Sequence II

TS II consists of early synrift sedimentary rocks comprised of strongly wedge-shaped sequences in basins larger than those of TS I, with lacustrine rocks that can be dominant in the basin's center. The TS II sequences exhibit evidence of deposition in more humid environments than those of TS I especially near the middle of their stratigraphic columns, as well as those of the younger sequences in the same basins. They are often separated by an angular unconformity from younger rocks in the up-dip portions of the fault blocks they comprise (Fig. 5). The age of the sequence is Anisian to Carnian [Middle to early Late Triassic]. The TS II is recognized in the Richmond, Taylorsville, Farmville, and associated basins of Virginia (Goodwin et al., 1986), the Newark basin, and the Fundy basin, and at least the Argana / HABC and associated basins of Morocco. The best-documented and type area is in the Richmond and Taylorsville basins of Virginia where TS II comprises most of the sequence preserved in the basins. In the Richmond and Taylorsville basin, TS II has the thickest and most widespread source rocks of the exposed early Mesozoic basins.

TS II overlies TS I unconformably in the Fundy (Stop 1.2), Argana basin, and HABC (Fig. 5). The outcropping strata of TS II in the Fundy basin comprises the Wolfville Formation (Fig. 4; Stops 1.2, 1.3, , 2.3, 3.2. 3.3), which is mostly fluvial. However, subsurface data (Withjack et al., 1995) suggest that there may be lacustrine equivalents to the Wolfville deeper within the basin as long suggested by Brown (in Wade et al., 1996) and are some seemingly cyclical lacustrine strata outcropping in the upper Wolfville Formation just north of Kingsport, Nova Scotia (Stop 3.5). In the Argana basin TS II consists of the Timesgadiwine Formation and consists of fluvial and lacustrine clastics with an especially well developed cyclical lacustrine sequence in the lower part of the Irohaline Member (T5) (Hofmann, et al., 2000) (Figs. 4, 8). Vertebrate paleontological data have long suggested that the lower Wolfville (Economy Member) is Anisian in age (Baird & Olsen, 1983) and the overlying middle Wolfville (lower Evangeline Member) is Carnian in age (Olsen et al., 1989), the latter age being supported by palynological associations from Fundy outcrops in New Brunswick (Stops 1.2 and 1.3) (Traverse, 1983; Nadon & Middleton, 1985; Wade et al., 1996; Fowell et al., 1996).

Figure 8: Cyclical lacustrine strata in TS II (Timesgadiwine Formation), Argana basin, near Rohala, Morocco. This unit is time equivalent to the Wolfville Formation in the Fundy basin.

In the HABC TS II is comprised of the Timenkar Group (El Arabi et al., 2006) (F3-5 of Biron & Courtinat, 1982) that includes most of the Oukaimeden Sandstones Formation. Based mostly on vertebrate remains augmented by some palynology (Biron & Courtinat, 1982 and Olsen, 1997), the age of TS II of the Fundy and Argana basins ranges from Middle to early Late Triassic age (Anisian – Carnian). In the HABC, El Arabi et al. (2006) have shown that the lower part of the Timenkar Group (lower unit of the “lower siltstone” = lower F4) is Anisian in age, while the base of the overlying Oukaimeden sandstones appears Carnian (Biron & Courtinat, 1982; Cousminer & Manspeizer, 1976); and in the Argana basin, the Aglegal Member of the lower Timesgadiwine Formation (TS II) produces apparent middle Triassic charophytes (Medina et al., 2003), consistent with a possible middle Triassic age for the capitisaurid amphibians from the same member (Olsen, 1997). Thus the Argana and HABC have TS II biostratigraphies parallel to what is seen in the Fundy basin. While TS II is very widespread in outcrop in the Fundy basin it has also apparently not been sampled in the subsurface in the Fundy basin.

Tectonostratigraphic Sequence III

Middle synrift sedimentary rocks of TS III form the bulk of the basin fill throughout eastern North America and Morocco, and they are also geographically the most widespread units (Fig. 2). In general, the sequence thickens less toward the border faults than any of the other tectonostratigraphic sequences and seems to represent the coalescence of many previously isolated basins of deposition. The age of the strata range from early Norian to late Rhaetian (early Late Triassic to latest Late Triassic) based on diverse lines of data including palynology and especially magnetostratigraphy (Cornet and Olsen, 1985; Fowell et al., 1996; Olsen, 1997; Muttoni et al., 2004). Thicknesses can exceed 6 km (Olsen et al, 1996a). There is often a basal, fluvial and fluvial-eolian interval followed by a thicker (>1 km) lacustrine sequence. In some basins the middle synrift sedimentary rocks are entirely fluvial. These are the “typical” red beds of northern Pangea. On the Scotian Shelf, the Grand Banks, and possibly Georges Bank, extensive Triassic-Jurassic evaporites are found in upper part of this sequence. The best-known portion of TS III is the Newark basin of New York, New Jersey, and Pennsylvania, considered the type area of the sequence.

An angular unconformity at least locally separates TS II and TS III in the Argana and Fundy basins. In some areas, at least of the Fundy basin, a profound angular unconformity (~90°) separates the Wolfville Formation of TS II from the Blomidon Formation (Fig. 4) of TS III. In other areas there may be a gentle angular unconformity, or none at all, but existing outcrops do not permit resolution of this uncertainty. In the Argana basin, the upper Timesgadiwine Formation is truncated by a gentle angular unconformity (<5°: Fig. 5) (Tixeront, 1973; Olsen, 1997), and then overlain by the Bigoudine Formation of TS III. TS III is entirely of Late Triassic age.

TS III is represented in HABC in Morocco by the lower part of the Rojdama Group (El Arabi et al., 2006) and includes basal thin eolian sandstones closely comparable to the eolian strata of the lower Bigoudine Formation (Tadrart Ouadou Member) of the Argana basin and overlying mostly red siltstones extremely similar to the upper Bigoudine Formation (Sidi Mansour and most of the Hasseine members). The eolian sandstones have traditionally been grouped in the Oukaimeden Sandstones, on purely grain size criteria, but genetically they are related to the overlying cyclical mudstones. Where exactly the boundary between TS II and TS III in the HABC is remains to be determined.

A characteristic facies in TS III of the Fundy and Moroccan basins is the sand patch fabric formed in saline basins (see Stop 2.3). The presence of sand patch fabric indicates the presence of sodium chloride brines at the sediment air interface producing halite efflorescent crusts. Halite pseudomorphs and abundant salt dissolution structures (Stops 2.3, 3.3) are also abundant in TS III strata in the Fundy and Moroccan basins, and in the Argana basin and HABC TS III ground water seepage in evaporative pans "salines" provide a local source of salt. TS III strata in the Essaouria contain thick halite sequences forming numerous salt structures, some of which produce hydrocarbons (the Meskala field, on-shore Morocco). The bulk of halite in the offshore Nova Scotia, Grand Banks, and Morocco almost certainly are hosted in TS III strata.

Tectonostratigraphic Sequence IV

Late synrift rocks, usually containing with an extensive sequence of tholeiitic lava flows of the Central Atlantic Magmatic Province (CAMP) and interbedded sedimentary strata, overlie TS III in many basins from the Culpeper basin north to the Grand Banks and Morocco (Fig, 2). Sedimentary rocks interbedded and overlying the basalts typically have much higher sedimentation rates than underlying sequences and, in exposed basins, have much better developed lacustrine high-stand lake deposits, suggesting more humid depositional environments or greater lacustrine basin depth. Tholeiitic lava flows with some thin interbedded sediments appear to represent CAMP and TS III in the southeastern United States. Traditionally, radiometric dates from the lava flows themselves have been difficult to interpret because of extensive alteration (see Sutter, 1988; Seidemann, 1989). However, U-Pb dates from gabbroid veins from the North Mountain Basalt of the Fundy basin and U-Pb and $^{40}Ar/^{39}Ar$ dates from a feeder intrusion of the Newark basin flows, and Newark basin flows themselves indicate dates of about 201–202 Ma for the flows (Hodych & Dunning, 1992; Ratcliffe, 1988; Sutter, 1988; Hames et al., 2000; Schoene et al., 2006). Milankovitch cyclostratigraphy sedimentary formations interbedded with and surrounding CAMP flows in the Deerfield, Hartford, Newark, and Culpeper basins is the basis for the estimate of the duration entire CAMP extrusion sequence of about 600 ky (Olsen et al, 1996b; 2003b), an estimate completely consistent with available radiometric dates (Hames et al, 2003; Marzoli et al., 2004). The duration of the CAMP within TS IV is thus similar to that of other flood basalt provinces, such as the Deccan and Siberian traps (Jaeger et al, 1989, Renne & Basu, 1991), and like the latter two LIPs, the CAMP has been implicated in the temporally associated mass extinction, at the Triassic-Jurassic boundary.

TS IV of the Fundy basin consists of the uppermost portions of the Blomidon Formation, and successively overlying North Mountain Basalt and McCoy Brook Formations. The homotaxial sequence in the Argana basin is the Bigoudine formation, succeeded by the Argana basalt and an unnamed unit designated T9 (Fig. 4). The equivalents in the HABC comprised the upper Rojdama Group and its eastern equivalents which is very similar to the Argana and Fundy sequences except the CAMP lavas are much more extensive and there is a complex of interflow sedimentary units, largely carbonates (Olsen et al., 2003b; Marzoli et al,, 2004; Whiteside et al. 2007, 2008).

The base of TS IV in the Fundy, Argana, and HABC basins includes the uppermost sedimentary sequence underlying the oldest CAMP lava flows. These sedimentary units include many last appearances of organisms marking the end-Triassic mass extinction event (Cornet and Olsen, 1985; Fowell & Olsen, 1995; Fowell & Traverse, 1995; Olsen et al., 2002a,b; Whiteside et al., 2007). In places where there is high-resolution data (i.e. the Newark basin), there is no evidence exists for an unconformity between TS III and IV (Fowell & Olsen, 1995). However, because half graben subside asymmetrically (Schlische, 1992) and TS IV is characterized by accelerated accumulation, increased subsidence rates, and half graben asymmetry, up-dip unconformities might be expected and in fact may have been observed locally in the Hartford basin (Olsen et al., 1992; Huber & Dutcher, 1999).

The uppermost Bigoudine and Blomidon formations (upper "White Water" and "Partridge Island members") and the Rojdama Group the begin with sand-patch-bearing red mudstones and sandstone, but within a few meters of the overlying basalts are a very distinctive suite of thin red, gray and black mudstone beds, which are generally palyniferous (barring local metamorphism from the basalt) that contain the palynological expression of the Triassic-Jurassic extinction event (Fowell & Traverse, 1995; Olsen et al., 2002a,b; Whiteside et al., 2007). The HABC has a precisely homotaxial pattern (Olsen et al., 2003b; Marzoli et al., 2004). In the Fundy basin, this extremely distinctive and mappable interval of the Blomidon Formation has been termed the Partridge Island Member (Olsen et al., 2003a). The lateral relationships of this unit are completely different than the underlying parts of the Blomidon Formation indicating, minimally, a reorganization of basin drainage system prior to the Triassic-Jurassic mass extinctions and the extrusion of the CAMP basalt flows.

The North Mountain Basalt and the Argana Basalt conformably overlie the uppermost Blomidon and Bigoudine formations in the Fundy and Argana basins, respectively (Fig. 2). Both basalt formations consists of at least two thick and several thin interbedded and overlying flows which have, mostly high-titanium, quartz-normative type tholeiitic chemistry (see summaries in Olsen et al., 2003b; Marzoli et al., 2004). The North Mountain Basalt has at least one interbedded sedimentary unit near its top (Fig. 4). In the HABC there is a more complex CAMP stratigraphy with at least two interbedded sedimentary units in the HTQ flows and an addition basalt flow sequence (the "recurrent basalt") of high-iron, high-titanium, quartz-normative type (HFTQ) tholeiitic composition. The HTQ basalts of the HABC are differentiated into three sub-types by "immobile" element chemistry (lower, intermediate, and upper basalt units) by Bertrand (Bertrand, 1991; Bertand et al., 1982) (Fig. 4). The relationships of these chemistries to Argana and Nova Scotia contingents of the CAMP will be discussed at stops on days 2 and 3 (Stops 2.1-2.3, 3.1).

The recent date of $^{206}Pb/^{238}U$ date of 201.27±0.03 Ma from the North Mountain Basalt gabbroids (Schoene et al., 2006) is indistinguishable (given interlab calibration uncertainties) from the $^{206}Pb/^{238}U$ date of 201.58 ± 0.28 from the marine Triassic-Jurassic boundary of Peru (Schaltegger et al., 2008). The ~201.5 date is almost certainly representative of the earliest CAMP flows throughout the province and indicates the initiation of eruption was temporally extremely close to the mass-extinction event. However, thus far there is no data that shows by superposition that the eruptions were synchronous or began before the extinction event (Whiteside et al., 2008).

Overlying the basalt flow formations in the Fundy basin and western and central Morocco are thin carbonate-dominated units succeeded by predominately red clastic rocks (Fig. 5). In the Fundy basin, the McCoy Brook Formation contains a basal carbonate-rich sequence of white, green, purple, and red lake level cycles (Scots Bay Member) (De Wet & Hubert, 1989; Tanner, 1996) and overlying predominately red clastic rocks, including some very local eolian sandstones (Hubert & Mertz, 1984). These produce relatively rich vertebrate assemblages (Olsen et al., 1987) that clearly post-date the extinction event (Olsen et al., 1987; 2002a). In the Argana basin, T9 (Hofmann et al., 2000; Tourani et al., 2000; Olsen et al., 2003a; Whiteside et al., 2007), contains a basal carbonate and black shale sequence, followed by predominately red and brown mudstones and sandstones with a few thin gray beds. Strata of TS IV are unconformably overlain by post-rift Quaternary strata in the Fundy basin and Early Jurassic conglomerates and marine strata in the Argana basin. In the HABC, post-CAMP sequences are highly variable, depended on post-rift erosion, but generally there is a conformable cyclical poorly known red-bed sequence that progressively has more evaporites and carbonates upward. This sequence is unconformably overlain by Jurassic to Cretaceous marine and marginal marine strata. The age of the youngest TS IV strata in all of these basins is poorly constrained but is probably at least as young as early Sinemurian in age, as is true further south in the Hartford basin (Kent & Olsen, 2008).

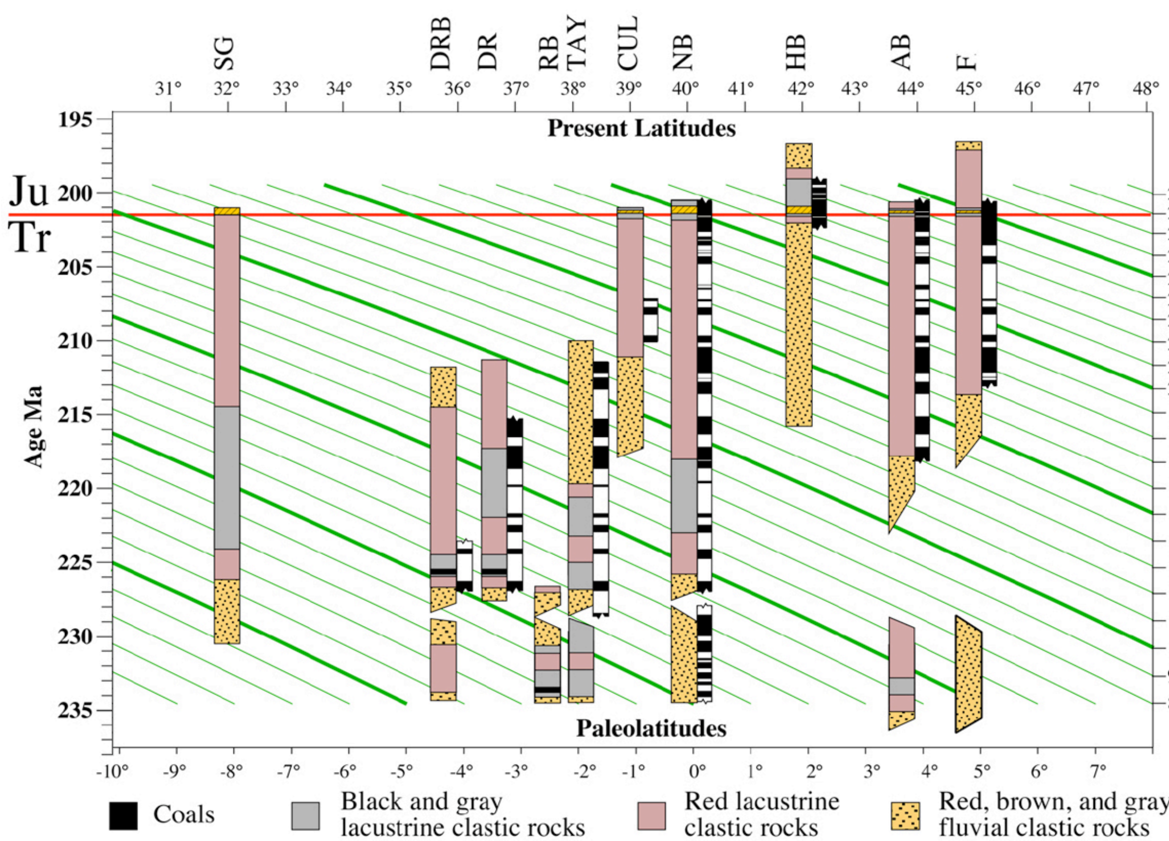


Figure 9: Time-geography nomogram showing the relationship between the main climate sensitive lithologies, age, geography, and latitude. Sections are correlated by magnetostratigraphy (black, normal; white, reverse). Slightly curved diagonal lines are lines of equal paleolatitude (modified from Olsen & Kent (2000) and Kent & Olsen (2000) using Kent & Tauxe (2005). Basins are: SG, South Georgia rift; DRB, Deep River basin; DR, Dan River basin; RB, Richmond basin; TAY, Taylorsville basin; CUL, Culpeper basin; NB, Newark basin; HB, Hartford basin; AB, Argana basin; F, Fundy basin, StA, St. Audrie's Bay, JL, Jamsonland basin. Ju/Tr is the end-Triassic extinction event.

Paleolatitudinal Context

The first order control on the geography of climate is latitude, via the effect on the angle of incidence of sunlight by the curvature of the Earth. Hence, the latitude at which a continent is located has the principle effect on climate, relative to other areas. While the broad pattern of the northward movement of the North American plate has been known for over 50 years, the details have become clear only recently as paleomagnetic data from the Newark Basin Coring Project (NBCP) became available (Kent et al., 1995; Kent & Olsen, 1999) and techniques were developed that could account for primary inclination error caused by sediment compaction (Tauxe & Kent, 2004; Kent & Tauxe, 2005). Based on these results, each of the basins in eastern North America and Morocco translated about 20° north during the Late Triassic and earliest Jurassic (Fig. 9). This means that the basins in eastern North America moved from the humid tropics to the arid subtropics during the period preserved in the synrift strata. Remembering that the other basins in eastern north America were also translating north and that there is about 10° of separation between the most northern and southern basins, the total latitudinal spread of the eastern North American rift records is over 30° (Fig. 8) and including the basins in Greenland and northern Europe, over 65° of paleolatitude. The sedimentary context of the Fundy and Moroccan basins needs to be viewed in light of this overarching latitudinal relationship because that is the first-order control on the development of potential sources rocks within the synrift strata as well as the nature and quality of potential sandstone reservoirs, and even synrift seals.

It is worth noting, that while it is commonplace to regard the Triassic as an arid time, there is in fact little justification for this. The notion of an arid Triassic, is largely rooted in the fact that major European and America cities lie in the Late Triassic Pangean subtropics. Modern paleomagnetic and paleoclimatic data show that the width of equatorial humid zone was apparently no different in the Late Triassic than today (Kent & Tauxe, 2005) and while today's polar areas are cold and dry, the Triassic polar areas were sites of extensive coal formation with no signs of ice. Thus, if anything, the Triassic continents were on average wetter, not drier, than today.

In conjunction with the changes in accommodation space caused by tectonics, it is the paleolatitudinal changes of the North American plate that were responsible for the first order vertical changes in facies seen in the basin sections. In particular, the Fundy and Moroccan basins, show an overall upward trend to more arid environments in younger strata from TS II through TS IV, as would be expected from drift from the tropics to subtropics.

Within the 65° of latitude spanned by the sequences discussed above there are profound changes in both the overall stratigraphic architectural style and the component sedimentary cycles of lacustrine strata. Olsen (1990) described three styles of lacustrine sequences in Eastern North America that, from the equator northward, are termed the Richmond, Newark, and Fundy type lacustrine sequences (Figs. 10, 11). Olsen & Kent (2000) added the Fleming Fjord and Kap Stewart style lacustrine sequences (Figs. 9, 10) for deposits further to the north. Carroll & Bohacs (1999) used the Olsen (1990) classification as part of their classification of lacustrine sequence typology.

Richmond style lacustrine sequences are characterized by indicators of long-term submergence and relatively muted climatic cycles (Fig. 9, 10). During relatively deeper water phases of the section, thick black shale sequences are interbedded with sandstones in the basin center, while much coarser strata interfinger along the basin margins. Black mudstones and carbonates are frequently microlaminated and contain a characteristic taphocoenesis of

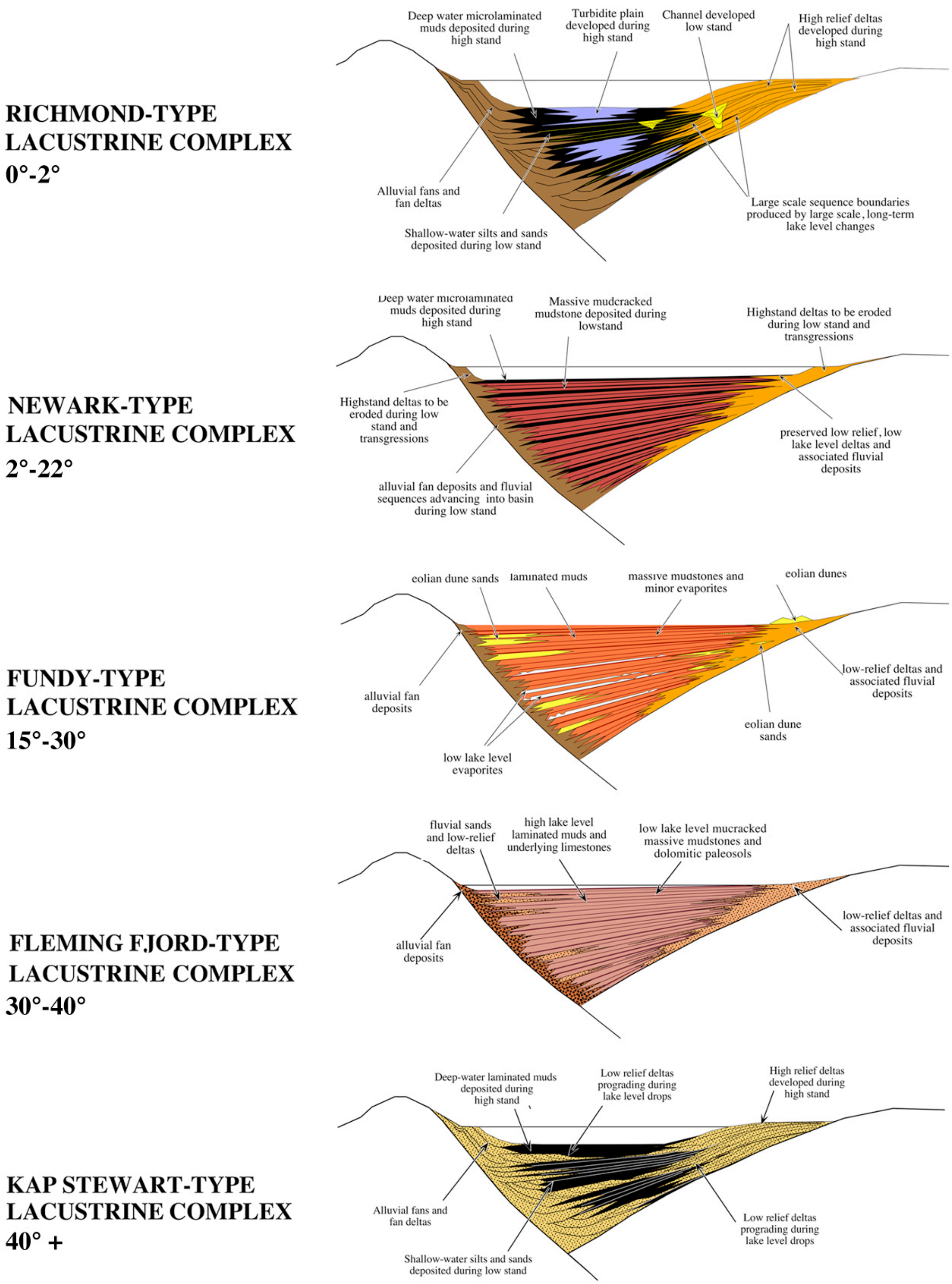


Figure 10: Idealized types of lacustrine facies complexes in the Triassic-Jurassic central Pangean Rifts. Note that no historical trend through the basin history is shown. Modified from Olsen (1990)

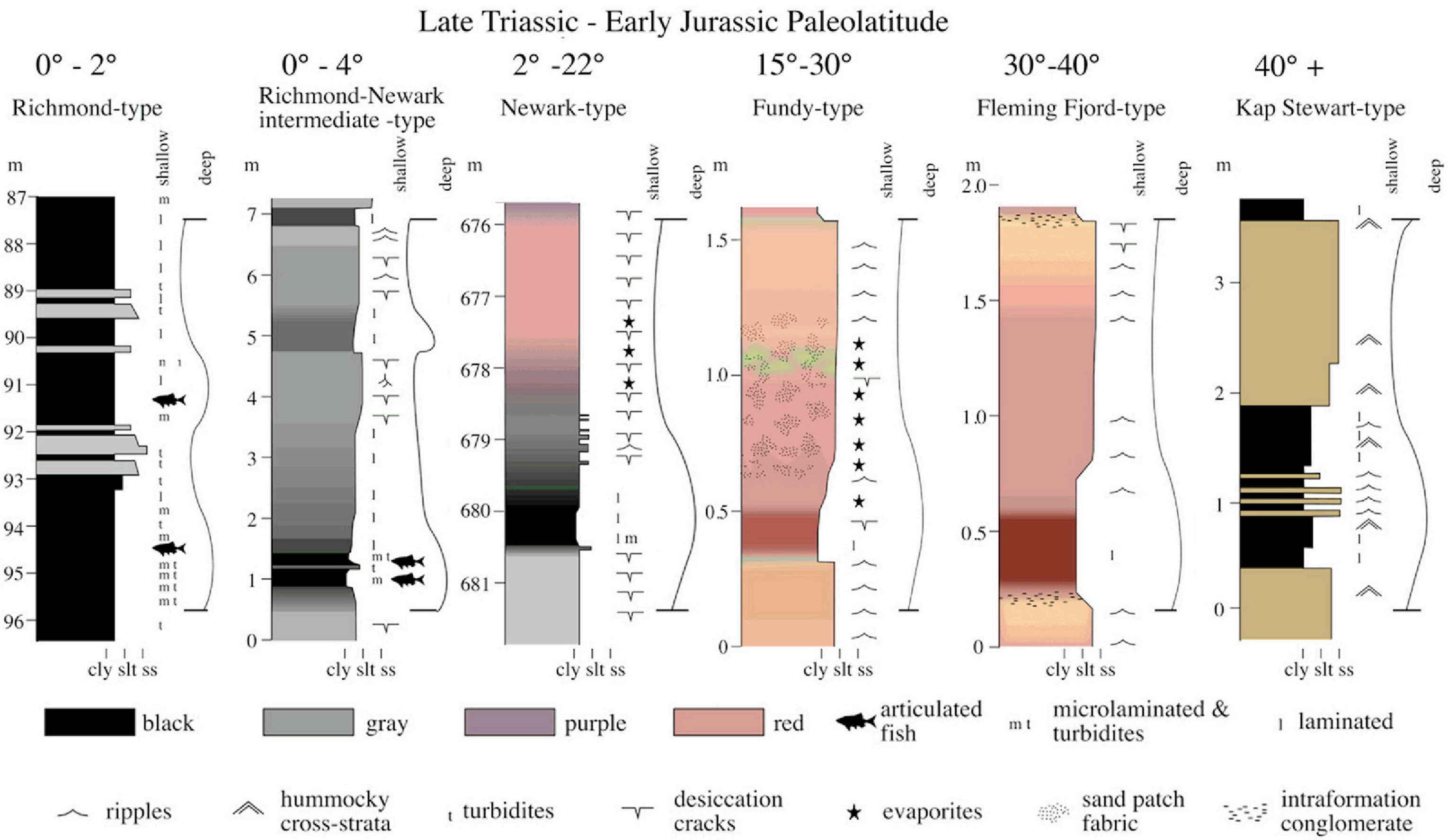


Figure 11: Representative actual section of individual ~20 ky Van Houten cycles in the different styles of lacustrine complexes. Modified from Olsen and Kent (2000).

articulated fish with an absence of bioturbation. Based on industry drilling logs and seismic data Cornet & Olsen (1990) suggested that there appears to be significant relief on fluvial and deltaic structures within and adjacent to the black shale sequences. Cores and outcrops indicate the presence of abundant turbidites, sublacustrine channels, and fan deltas far into the basin. Shallower water phases of the basin sections have more prominent cyclicity, but the drier parts of the cycles tend to be sandy with intense bioturbation. Significant coal bearing intervals can be present in the shallower water intervals (e.g. the "productive coal measures"). Richmond style lacustrine sequences have been found in the Richmond, Taylorsville, Briery Creek, and Scottsburg basins of Virginia and may be subsurface in the Newark and Gettysburg basins.

Newark-style lacustrine sequences are the most common type in eastern North America (Figures 10, 11) and have a very extensive descriptive literature (Smoot & Olsen, 1994; Olsen, 1997), being characterized by very pronounced cyclicity caused by alternating intervals of submergence and desiccation at time scales of 20 ky. Basin-center facies tend to be fine-grained with low-stand intervals dominated by red or gray mud-cracked massive mudstone and rooted mudstone. Deeper water intervals consist of black to red shales, with black shales frequently being microlaminated with the same basic taphoceoensis as in the Richmond-style sequences. Coarse clastic rocks are more restricted to the margins of the basin than in Richmond-style sequences. Deeper-water parts of the basin sequences can be mostly gray, but each sedimentary cycle has evidence of desiccation, even though the cycles may have a very prominent black shale. Shallower water phases of the basin sections tend to be mostly red clastics with less bioturbation than Richmond-style sequences. Coals are rarely present, but some are present in the more southern basins. Evaporites can be fairly abundant, although no pure bedded evaporites have been identified. The evaporite-bearing beds are almost always in the shallower-water parts of cycles, usually with desiccation cracks and sometimes reptile tracks. The overall stratigraphic architecture tends to be characterized by virtually no evidence of significant topography. Newark-style sequences have been found in the Deep River, Dan River, Taylorsville, Culpeper, Gettysburg, Newark. Hartford, Deerfield, and Argana basins. Those of the Deep River, Dan River, Taylorsville basins are transitional to Richmond-style sequences, while older strata of the Argana and HABC basins are transitional between Newark- and Fundy-style sequences.

Fundy-type lacustrine sequences are characterized by evidence of very low synsedimentary topography and are highly cyclical, as in Newark-type sequences (Figs. 9, 10) and comprise most of what we will see in the field. However, Fundy-type sequences tend to lack black and gray shales, and microlaminated black shales are completely absent. Instead, such sequences are dominated by red massive mudstones and eolianites, with abundant evaporites, which can include massive beds of sulfates and halite. The most common facies is the sand-patch massive mudstone, which is a muddy eolianite that accumulated on an efflorescent evaporitic crust (Smoot & Olsen, 1988). See Days 2 and 4, below. Thick and widespread eolian dune sequences are also present (Hubert & Mertz, 1980). Significant bodies of coarse clastics tend to be limited to the margins of the basins, although abundant, thin tabular "sheet flood” sandstones extend far into the basin. Fundy-type lacustrine sequences have been identified for certain in the Fundy, Argana, and HABC basins, but are probably more widespread, most likely occurring in other basins in Morocco, the Scotian Shelf, Grand Banks, Iberia, and in older Triassic strata in Europe, although we have not confirmed their presence through first-hand observation. These types of sequences exhibit salt dissolution features in outcrop and are often interbedded with the largest evaporite sequences in the Essaouria basin and other basins offshore Morocco and the

Scotian shelf (Olsen et al., 2000, 2003b). This is probably also the case for the salt sequences of the Grand Banks.

Fleming Fjord-style lacustrine sequences are similar to Fundy-type sequences in lacking synsedimentary relief and in being dominantly red-bed sequences (Figs. 9, 10). However, mudcracked massive mudstones and bioturbation (both by invertebrates and roots) are more common, while evaporites and eolianites are rarer, although occasionally present, especially in more southern examples. Thin, stromatolitic carbonates, dolostones, and ripple-laminated sandstones are also present along with rare, thin coals. This type of sequence is characterized by common evidence of liquefaction, which apparently mired prosauropod dinosaurs (Sander, 1992; Jenkins et al., 1994). The best-described examples of this type of Fleming Fjord-style lacustrine sequence are the Trossingen (Knollenmergel) and Arnstadt (Steinmergelkeuper) formations of the Germanic basin (Seegis, 1993; Bachmann et al., 1998, Beutler, 1998; Reinhardt & Ricken, 1998; Kellner, 1998) and the Fleming Fjord Formation of the Jameson Land basin in east Greenland (Clemmensen, 1980). However, Fleming Fjord-style lacustrine sequences are almost certainly more widespread than this list would indicate and probably characterize portions of most of the Triassic basins of northern Europe. Strata seemingly transitional between Fundy and Fleming Fjord-style lacustrine sequences are evidently present in the rift basins of Great Britain, such as the Mercia Mudstone Group (Kemp & Coe, 2007), that contain significant bedded evaporites at depth.

Kap Stewart-type lacustrine sequences resemble Richmond style sequences in the abundance of sand and evidence of higher-relief structures such as deltas (Figures. 10, 11). However, aqueous wind-dominated sediment reworking is much more prevalent (Dam & Surlyk, 1992) and coals are much more abundant than in Fleming Fjord-style sequences. While black shales are very common, microlaminated units, with their characteristic taphocoenesis have not been specifically described. This type of sequence is apparently very common in Northern Europe and East Greenland having been described from the Danish-Polish basin of Denmark, Sweden, and Poland (Pienkowski, 1984; Norling et al., 1993), the Haltenbanken area (Hollander, 1984), and East Greenland (Dam & Surlyk, 1992, 1993). It is not known how far north the Kap-Stewart type sequences extend. We note that the presence of coals at high latitudes is to be expected under the high precipitation regimes successively predicted by some global climate models for a "hot house" Triassic world (Kutzbach, 1994). We also note that some portions of Kap Stewart-type sequences in Sweden, for example, contain some beds with marine invertebrates (Germanic and Danish-Polish basins), indicating at least some incursions of marine waters. Kap Stewart-type sequences were deposited from about 45° N paleolatitude to the north. They give lie to concept of the “arid Triassic”.

These lacustrine sequences generally characterize very large portions of the basin sequences and different types of sequences can follow each other in vertical succession. The vertical succession of lacustrine types follows the general climatic pattern previously outlined. Richmond type sequences developed in the humid equatorial tropics: Newark type formed in the seasonally and millennially arid transition zone into the arid tropics; Fundy-type sequences were deposited within the arid tropics; Fleming Fjord-type sequences formed in the transition to the temperate zone; and Kap Stewart-type sequences formed in the humid temperate zone under "hot house" conditions. The latitudinal arrangement of lacustrine types basically reflects the large-scale zones of differing precipitation/evaporation ratios. As the basins translated north with central Pangea, the major lacustrine sequence type changed with the passage into different climate zones. Examples in which one sequence type unambiguously follows another in vertical sequence within a single basin include the Taylorsville basin of Virginia (Richmond- up into Newark- type; LeTourneau, 1999), the Argana basin, Morocco (Newark- up into Fundy- type: Olsen, 1997), and the Jamesonland basin, Greenland (Fundy- up through Fleming Fjord- and Kap Stewart-type).

Sedimentary Cyclicity

Perhaps the most striking aspect of the synrift sequences on the conjugate margins of North America and Africa is the prevalence of cyclicity in lacustrine lithologies evident at many thickness scales (Fig. 12). This sedimentary cyclicity was primarily controlled by lake-level variations following quasi-periodic Milankovitch-type climate cycles governed by celestial

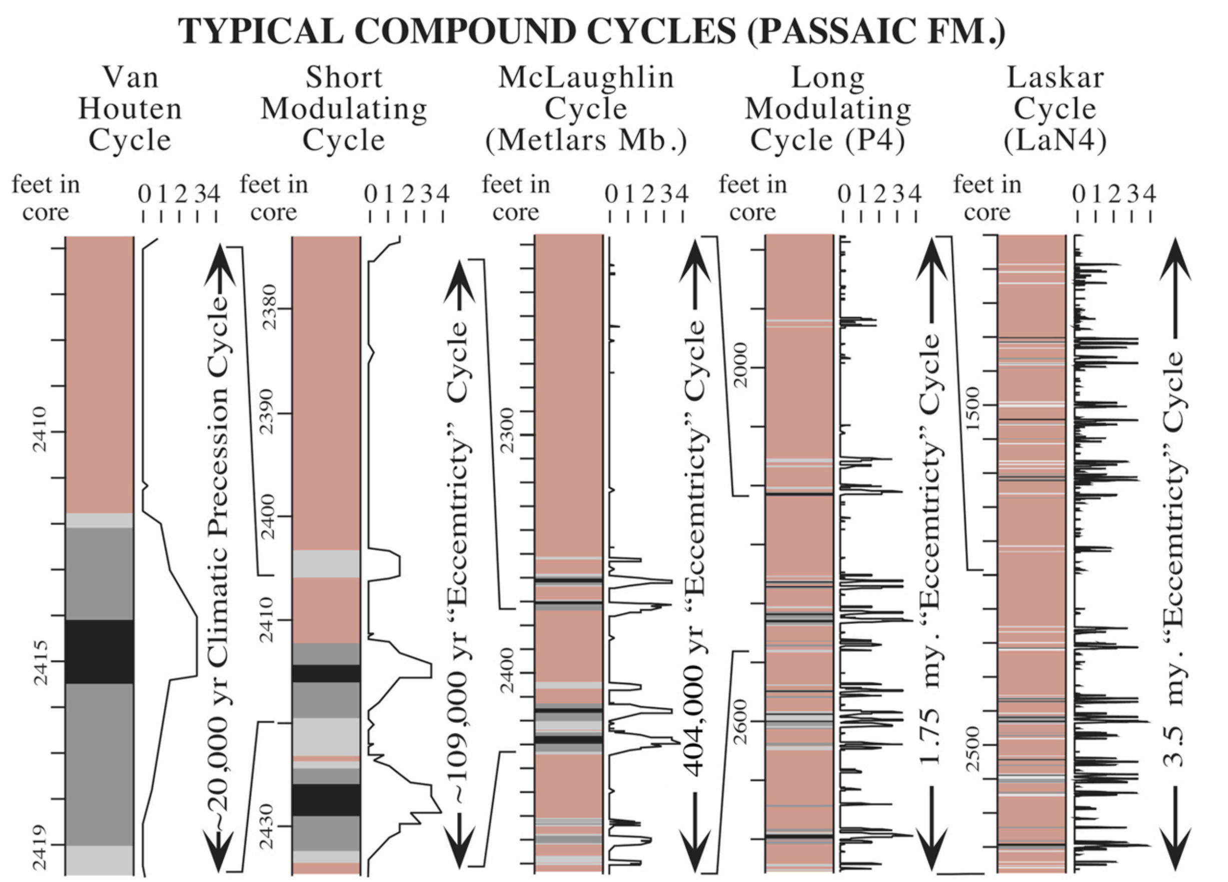


Figure 12: Hierarchy of astronomically forced lacustrine cycles in CAM basins. This is an example drawn from the Newark Basin coring project cores of the Passaic Formaton, a Newark-style lacustrine complex. Modified from Olsen & Kent (1999).

mechanics (Hays et al., 1976). They therefore control the partitioning of source rock within the basin sections, within their overall climatic context. The careful and detailed review by Smoot (1991) explains the climatic significance of various sedimentary facies and cycles within these basins. Although cyclicity is pervasive, the best-documented example of cyclicity and

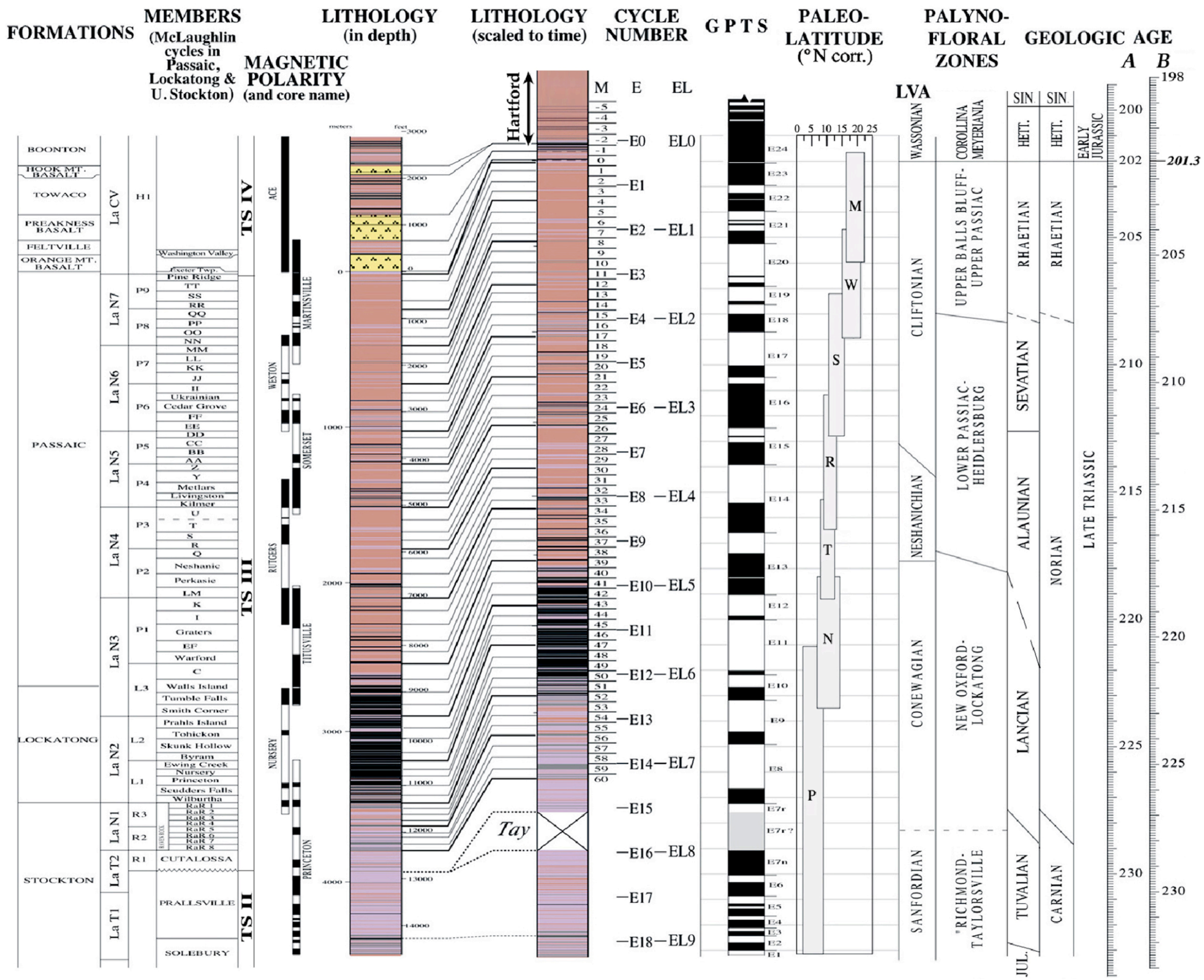


Figure 13: The Newark Basin Astronomically-calibrated Geomanetic Polarity Time Scale (NBAGPTS),

periodicity is from the Newark basin (Fig. 13). Sedimentary cyclicity in the lacustrine portions of the basin was first described by McLaughlin (1933), who demonstrated the great lateral extent of the large-scale (~100 m) alternations of gray to black shales and red mudstones (McLaughlin, 1933, 1946, 1959). Van Houten (1962, 1964, 1969, 1980) extensively studied the Lockatong Formation and provided a detailed look at Newark basin cyclicity, ascribing it to climatic variations controlled by celestial mechanics. Olsen (1986), following the paradigm of Hays et al (1976) for Quaternary cyclicity, used Fourier analysis to obtain quantitative assessments of the periodicity of the cyclically bedded outcrops, thus corroborating Van Houten's hypothesis (Fig. 14). Pervasive cyclicity in most of the Newark basin section was confirmed by continuous coring of the Newark basin by the Newark Basin Coring Project (NBCP) (Olsen et al., 1996a; Olsen & Kent, 1996) and the Army Corps of Engineers (ACE) (Fedosh & Smoot, 1988; Olsen et al., 1996b). Together, over 12 km of core from these projects cover virtually all of the approximately 7 km–thick Newark basin section, including more than 5 km of cyclical lacustrine rocks that show patterns consistent with the Milankovitch theory of climate change (see Smoot & Olsen, 1994; Olsen & Kent, 1996, 1999) (Fig. 13). Subsequent evolutive Fourier and wavelet analysis (Fig. 15) coupled with new independent radiometric dates confirm the Milankovitch origin of the cyclcity (e.g., Olsen & Kent, 1999; Furin et al., 2006). Together with the ACE cores and a recently completed cyclostratigraphy and magnetostratigraphy for the Hartford basin (Kent & Olsen, 2008), the NBCP core form the basis for the Newark Basin Astronomically Calibrated Geomagnetic Polarity Time Scale or NBAGPTS (Fig. 13), which will serve as the geochronological foundation for most of the quantitative statements made in this guidebook.

As seen in the cores and outcrops of the Newark basin lacustrine units, as well as in several other CAM basins, including the Fundy and Moroccan basins, the fundamental sedimentary cycle is a transgressive-regressive lacustrine sequence termed a Van Houten cycle (Olsen, 1986, 1990) (Fig. 12). Vertical sequences of these cycles trace out three orders of modulating cycles; these cycles are expressed as variations in the degree of development of sedimentary structures indicative of subaerial exposure or deposition in deep water within successive Van Houten cycles. The short modulating cycle contains 4–6 Van Houten cycles; the McLaughlin cycle exhibits 4 short modulating cycles; and the long modulating cycle is made up of about 5 McLaughlin cycles (Fig. 12). There is even a longer modulating cycle comprised of two successive long modulating cycle (Olsen & Kent, 1999). Frequency analysis of sections exhibiting this cyclicity by Fourier and moving window ("evolutive") techniques (e.g., Fig.13) indicates that the cyclicity conforms to the expectations of Milankovitch climate theory for the tropics. Specifically, the climate changes that controlled lake level were governed by the cycle of climatic precession, modulated by the 100 ky, 413 ky, 1.75 m.y. and 3.5 m.y. so-called eccentricity cycles of the Earth's orbit.

As expected from Milankovitch theory, the expression of the insolation cycles in continental climate, and hence the rock record, varies dramatically with latitude in the CAM basins. Within specific time intervals, the distinctive frequency characteristics of the cyclicity as well as the overall facies (i.e. humid vs arid) change in a methodical way that is well explained in the Milankovitch paradigm. In theory, the region around the equator should be dominated by a "double" climatic precession cycle of 10 ky that gives way farther from the equator to a 20-ky cycle of the type seen in the Newark basin and commonly associated with the climatic precession cycle (Crowley et al., 1992). In addition, the middle and high latitudes should be dominated by the effects of the 41-ky cycle of the obliquity of the Earth's axis. In fact, this domination is manifest in the eastern North American basins. This pattern of latitudinally dependent cyclicity is apparently in very good agreement with the paleogeographic reconstruction (Kent & Tauxe, 2005), which places Virginia at the equator in the early Late Triassic (Fig. 8). The change in cyclical mode is paralleled by a dramatic change in the dominant sediment type, with coals and black shales in the south giving place to eolianites in the northern Newark Supergroup basins and Morocco (Olsen et al. 1989). The cyclicities so well displayed in the these basins provide not only a metronome for calibrating other basin phenomena, such as the magnetic polarity sequence from the basins, accumulation rates, and hence indirectly subsidence, but also were the first clear demonstration of the local effects of Milankovitch cycles on continental climate over a broad latitudinal swath (Fig. 11). Recent coring in the modern African Great Lakes seems poised to show a similar pattern (e.g., Cohen et al., 2007).

However, the Fundy basin and basins in Morocco are dominated by sand patch cycles of Fundy-type lacustrine sequences , and these cycles tend to be thin (~1 m), and in contrast to Newark type sequences, exhibit considerable noise in their spectral properties (Stop 4.2). Nonetheless, with time control provided by magnetostratigraphy, it is possible to not only identify the Milankovitch pattern, but even to identify individual cycles and sequences of cycles in the time (thickness) domain (Fig. 16), but also in the frequency domain in wavelet spectra both in sedimentary facies and in geophysical properties (Fig. 17) allowing direct comparisons with the Newark basin sequences. The correlative sequences display a remarkable amount of order, bespeaking a strong climatic and tectonic control that overrides local effects.

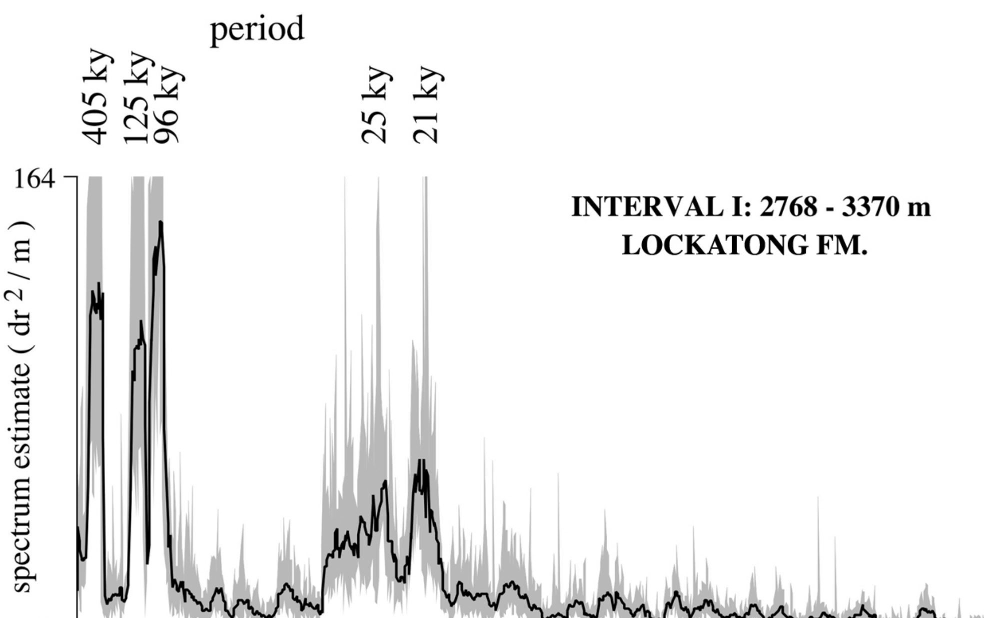


Figure 14: Power spectrum of lacustrine sedimentary facies (depth ranks) from the Nursery and Titusville NBCP cores. The spectrum is on an untuned data series with the low frequency peak, corresponding to a McLaughlin cycle, set to 405 ky (modified from Olsen and Kent, 1996.

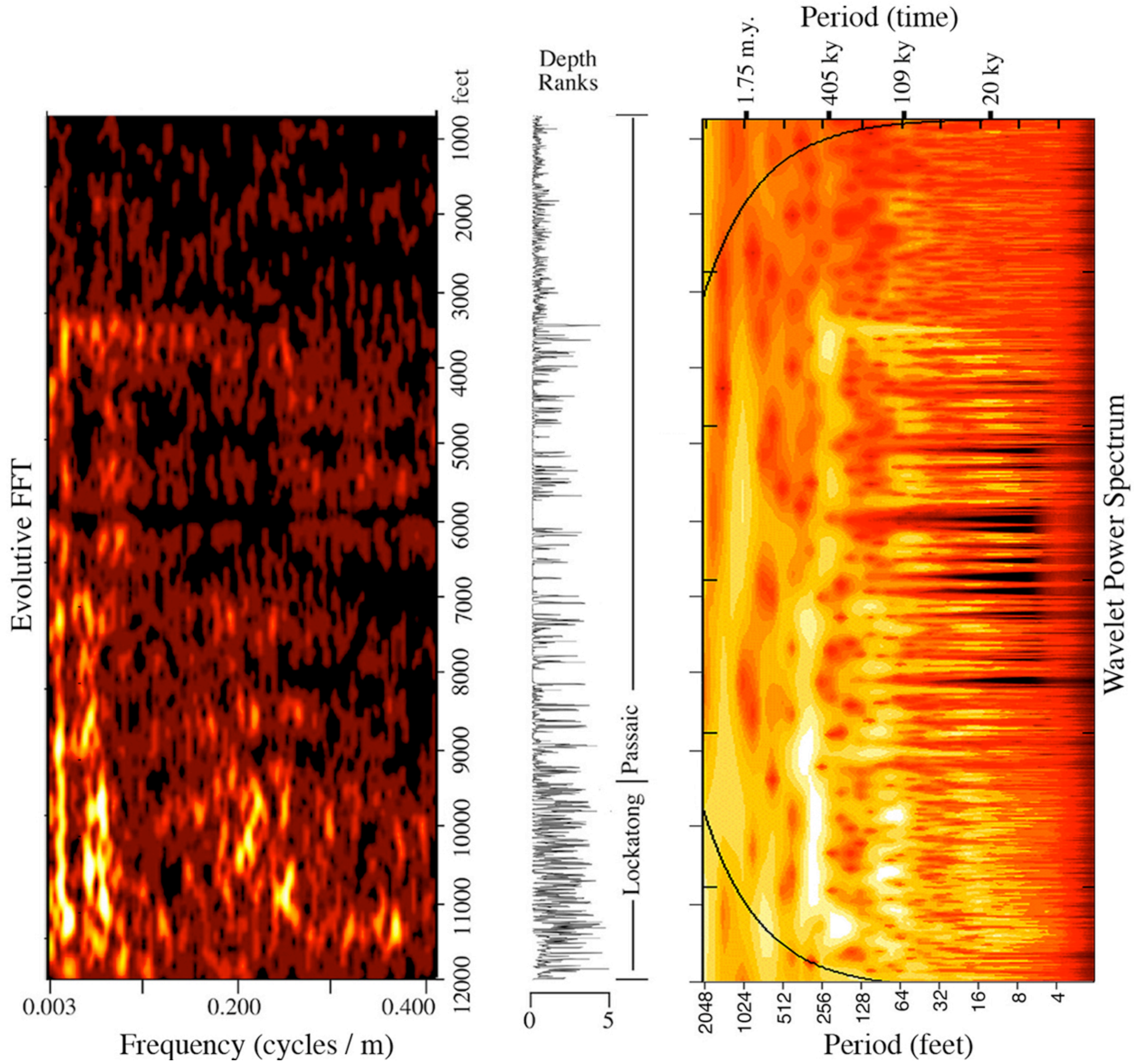


Figure 15: Evolutive FFT (left) and Wavelet spectrum (right) of untuned lake level (depth rank) data. High power is indicated by brighter colors. Note that low and high frequencies (and periods) move in tandem according to accumulation rate changes so that the ratios of short and long periods in thicknesses are constant. This is especially obvious in the Wavelet spectrum. Evolutive FFT is from Olsen & Kent (1999).

Biotic/Climatic Events and Evolutionary History

Although much of the variability in climate-sensitive rocks is compatible within the framework of northerly drift, some of the variability is not. Stated more specifically, just before the Triassic-Jurassic extinction event, the frequency of more humid facies increases dramatically across all latitudes regardless of tectonic setting along this more or less meridional tract. Yet, North America should be drifting farther into the arid belt during this time. The direction of apparent climate change is thus at variance with the predictions of drift. In addition, this very rapid change in apparent climate came without a change in the rate of drift. There is no sign of an abrupt increase in wetness in the overall drying trend seen in the early Mesozoic deposits of the western United States, and thus the change does not seem to be global. One simple explanation is that the climate became more humid regionally, although within these basins some of the trend could be due to increased extension marking the base of TS IV. It is interesting to speculate whether this humidification could have been associated with the Rhaetian–Early Jurassic transgression, marked in the northern part of the rift province by the appearance of very widespread marine-derived evaporites. On the other hand, the Triassic-Jurassic boundary itself was one of the largest mass extinctions of all time (Hallam, 1992; Olsen et al., 1988; Sepkoski, 1993). The conjugate margin rift basins are one of the principle and highest resolution recorders of the event (Olsen et al., 1988; Olsen et al., 1990; Silvestri & Szajna, 1993; Fowell & Olsen, 1995; Fowell et al, 1994). Explanations for the mass extinction run from the nearly contemporaneous flood volcanism (Courtillot, 1994), to marine transgressive-regressive couplets (Hallam 1992), to asteroid or comet impact (Dietz, 1986, Bice et al., 1992; Olsen et al., 1988, 2002a, b). The climate change evident near the end-Triassic extinction event in the CAM basins, as well as the flood basalt event that just postdates it, could be part of the causal chain. This climate change may have involved large, but temporary increases in CO_2 that may have profoundly effected primary productivity of the oceans and land both directly and indirectly through weathering-driven fertilization, and the extinction event itself may have driven changes in the biological pump in lakes as well as in the ocean.

A major factor in source rock development is the overall evolutionary context of the system being investigated. Despite the seemingly "primitive" nature of the organisms involved there are major differences in the major organisms that were important primary producers in the Mesozoic, especially during its early half. Diatoms, for example, are the most important phytoplankton in presently productive water bodies, but they are absent from early Paleozoic and virtually absent from early Mesozoic marine and non-marine strata. Instead, cyanobacteria were among the main components of the phytoplankton of lakes together with dinoflagellates and prasinophytes (photosynthetic but mixtrophic flagellate "green algae") in the oceans. The absence of major taxa, especially diatoms, from lacustrine and marine systems limits the degree to which ecological analogies can be expected to be useful in ancient systems. The evolution of new taxa with key ecological and physiological innovations changed major aspects of the production of source rocks in ways that are only now being realized. As we examine various localities in the field, we will point out some of the implication of these innovations.

PART 2: Fieldguide to the Fundy Rift Basins and Comparison to the Moroccan Conjugate Margin Basins

Introduction

Geography and tides dictate the arrangement and schedule of field stops for this fieldtrip and guidebook. The fieldtrip is correspondingly divided into three areas around the coast of the Bay of Fundy examined over 4 days (Fig. 18) corresponding to the western border fault and foot wall area in New Brunswick (Day 1), the northern strike-slip Minas fault zone (Day 2), and the hanging wall or hinge margin (Days 3 & 4). At most stops, we are captive to the tides and it is possible because the immutable time constraints posed by the tides, a stop might have to be skipped if we linger at a particularly interesting spot or the weather is uncooperative, making intertidal rocks deadly slippery. However, given these constraints, we should have a good overview of Fundy basin stratigraphy, facies, and structure providing a natural basis for comparison to the Scotian Shelf and Moroccan basins on their conjugate margins.

Figure 18: Field stops for this trip. (Derived from ©2008 Google Map Data, ©2008 Navteq™ .)

Field Day 1: Western Fault-Bound Side of the Fundy Basin

The main boundary fault system of the northern half of the Fundy basin lies along the northwestern edge of the Bay of Fundy, in New Brunswick, Canada, the faults with the most observable displacement lying just offshore (Withjack et al., 1995). Outcrops along the northwestern side of the Bay of Fundy reveal the not only parts of this boundary fault system in footwall rocks, but also several inliers, on rider blocks of this fault system that give us glimpses of basin facies not seen on the strike slip margin or on the hanging wall side of the basin (**Stops 1.1-1.3**). The pre-rift footwall rocks are important, not only structurally, but also because they contain organic rich Paleozoic strata that source the only producing onshore gas field in Eastern Canada, the McCully Field (Stop 1.4), and the only productive (formerly) oil field, the Stony Creek Field (Fig. 18). Some of these strata underlie portions of the Fundy rift offshore in the Bay of Fundy and hence could source syrift reservoirs. In this leg of the field trip we will look at three stops in synrift strata (Stops 1.1-1.3) and one stop in organic-rich prerift strata (**Stop 1.4**), proceeding basically southwest to northeast.

<u>Begin Fieldtrip:</u> 1 Market Square, Saint John, New Brunswick, Canada E2L 4Z6 (St. John Hilton) (about 1:04 hours to Stop 1.1).

Go east on Union St. toward Provincial Secondary Route 100	0.00 km
Turn left onto Smythe St/ Provincial Secondary Route 100.	
Continue to follow Provincial Secondary Route 100.	0.83 km
Turn slight right	0.05 km
Turn slight right	0.10 km
Turn right onto Somerset St.	0.23 km
Take the hwy 1 E/Route-1 E ramp toward Sussex	0.26 km
Merge onto HWY-1 E	11.8 km
Take exit 137A to merge onto HWY-111	
toward Saint John Airport/St Martins	8.8 km
Turn left at HWY-111/Loch Lomond Rd	
Continue to follow HWY-111	29.7 km
Turn right at West Quaco Rd	3.3 km
Turn left	0.5 km
End of road: Stop 1.1, Leave vehicles.	

Walk down path on south side of road to the ladder down to the beach (get permission). Then proceed to the southeast ~0.8 km to 45°18.638', N 65°33.666'W where we will begin our transect. We retrace our path northeast to the ladder and if the beach ridges to the northeast are impassible, walk back up the ladder to the road and then walk east down path on north side of the road to the beach and proceed to the north end of the transect at 45°19.009'N, 65°33.240'W.

Stop 1.1: Honeycomb Point, West Quaco, St. Martins New Brunswick

Transect begins at 45°18.638', N 65°33.666'W, ends at 45°19.009'N, 65°33.240'W

Main Points:

Oldest known Fundy rift basin strata of TS I of probable Late Permian age.
Very similar to non-rift Cap Aux Meules of the Magdelan basin.
Arid climate, sand seas contrasting with Late Permian deposits in Morocco.
Excellent potential reservoirs on possible source rocks (elsewhere).

This is the type area of Magnusson's (1955) Honeycomb Point (raised to formation status by Klein, 1962), consisting of red and brown clastic rocks of lacustrine, fluvial, and eolian origin. At the southwestern end of the outcrop belt, at the beginning of our transect, is a spectacular triple unconformity mapped by Plint & van de Poll (1984) (Fig. 19).

A steeply dipping unconformity between Precambrian Coldbrook Group metabasalts overlain a thin layer (5-10 m) of marine, Mississippian Windsor Group limestone, itself disconformably overlain by near vertical alluvial conglomerates of possible Mississippian Hopewell Group. This is all juxtaposed by a profound unconformity with the gently dipping basal Honeycomb Point Formation (Fig. 20).

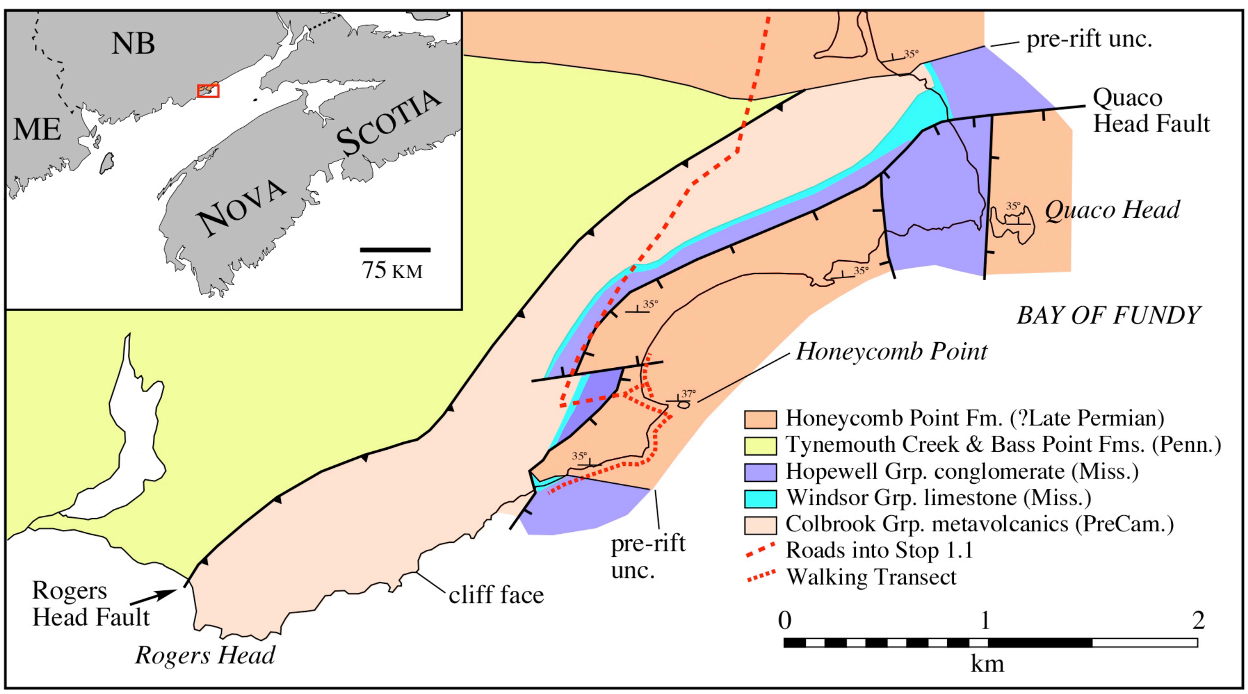


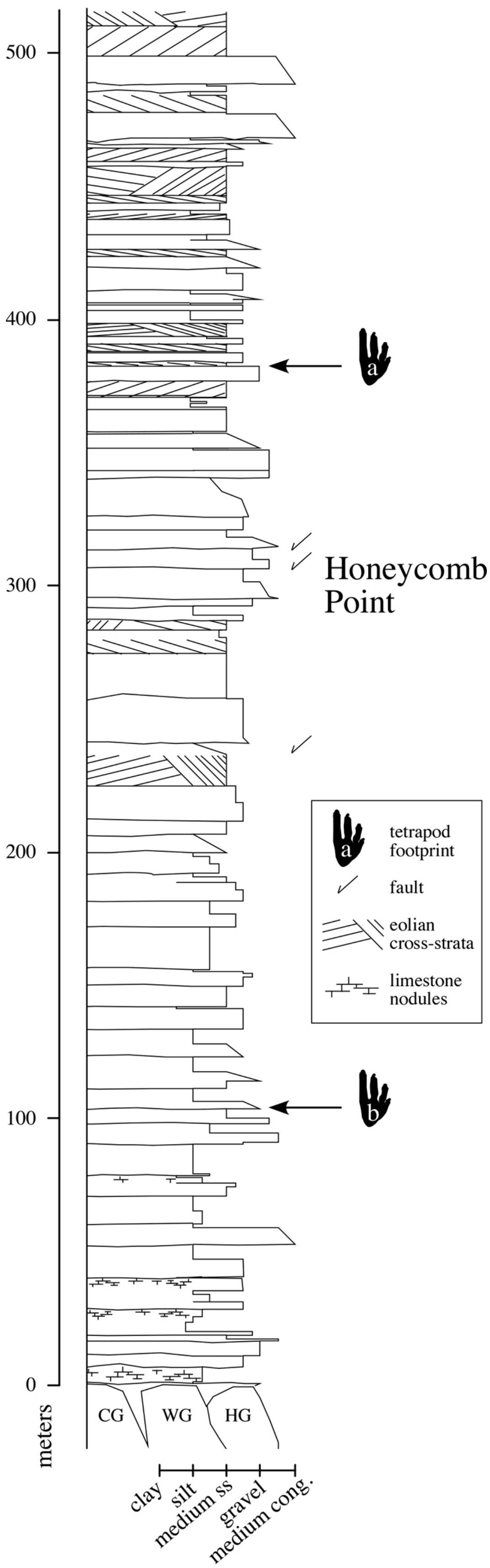
Figure 19: Geological map of the Honeycomb Point area, Stop 1.1. Adapted from & van de Poll (1984). Red box in inset shows position of the map.

The Honeycomb Point Formation at this section (Browns Beach Member of Nadon & Middleton, 1985) consists of a red basal carbonate rich pedogenic interval (caliche) followed by red fluvio-lacustrine strata, which is in turn followed by eolian strata (Figs. 20, 21) spanning over 500 m. There is little direct age information for this unit. There are various invertebrate and

Figure 20: Triple Unconformity at Stop 1.1 with Honeycomb Point Formation unconformably overlapping Paleozoic and Precambrian rocks.

plant ichnofossils, but the forms present are not age-diagnostic. There are uncommon tetrapod footprints (Fig. 22); these are mainly of interest because none of the forms found so far look at all like typical forms from undoubted Triassic strata of eastern North America, and although there is a passing resemblance to forms from the Late Permian of Morocco (c.f., Hmich et al., 2006), none are well enough preserved to be age-diagnostic. There are no palynomorphs, macrofossil plants (excluding roots), vertebrate bones, or invertebrate body fossils. Most relevant are as yet unpublished paleomagnetic polarity and pole position data that indicate nearly entirely reverse polarity and Late Permian poles (P.M. Letourneau and D.V. Kent, pers comm.), and thus a Late Permian age is consistent with all available information.

Interestingly, the Honeycomb Point Formation finds a close lithological counterpart in the Cap Aux Meules Formation of the Magdalen basin exposed along the flanks of salt diapirs in the Gulf of St. Lawrence (Fig. 23). The Cap Aux Meules Formation (Brisbois, 1981; Tanczyk, 1988) consists largely of eolian sandstone and interbedded fluvial deposits disconformably overlying Early Permian age strata. The strata appear Late Permian in age based on paleomagnetics (Tanczyk, 1988) in the absence of other data. The geometry of the Cap Aux Meules Formation is poorly constrained, but like other late Paleozoic strata of the Magdelenic basin the formation is probably relatively thin, covering a broad area, very unlike a wedge. In contrast, while exhibiting similar facies and apparent age, the Honeycomb Point Formation is minimally 900 m in thickness and its correlative, the Lepreau Formation to the southwest, is over 2 km in thickness, demanding a different style of growth of accommodation space than seen in the Magdalen basin. While definitive features of growth against a border fault, such as conglomerates coarsening towards the footwall or a fanning geometry are lacking because of the very small area of the preserved strata, the thickness of the Honeycomb Point and Lepreau formations are consistent with deposition in a half graben complex.



The presence of abundant eolian sandstones and well developed caliche is consistent with a sometimes very arid environment. The coeval Ikakern Formation (TS I) of the Argana basin (Fig. 23) and the Cham-El-Houa beds of the HABC of Morocco lack eolian sandstones and thick caliche and seem to have more abundant shallow-water aquatic deposits locally with abundant tetrapod footprints (Hmich et al., 2006), as well as tetrapod remains including the amphibian *Diplocaulus* (Dutuit, 1988) and herbivorous reptiles (pareiasaurs: Jalil & Janvier, 2005). This is consistent with the Argana basins being significantly to the south and hence in more humid climes during the Late Permian due to central Pangea being rotated clockwise compared Late Triassic coordinates (Fig. 7).

The eolian and some of the fluvial strata of the Honeycomb Point Formation would seem to make excellent reservoirs, although actual measurements have not been published. Were they to be juxtaposed against organic-rich prerift strata of the Moncton basin, which could be the case below Chignetco Bay, they could provide a mechanism for significant accumulation of hydrocarbons.

Return to vehicles. (about 10 min. to Stop 1.2)

Drive west
Turn right toward West Quaco Road
Continue straight onto West Quaco Road
Turn right at Fundy Scenic Route/HWY-111
Turn right at Main Street (becomes Big Salmon River Road) 3.2 km
Turn right and stop at beach on right just to the northeast of the beachfront restaurants.

Figure 21: Measured section at Stop 1.1: CG, Coldbrook Grp. metavolcanics; WG, Windsor Grp. limestone; HG, Holbrook Grp. Conglomerate. Letters in footprints refer to Fig. 22.

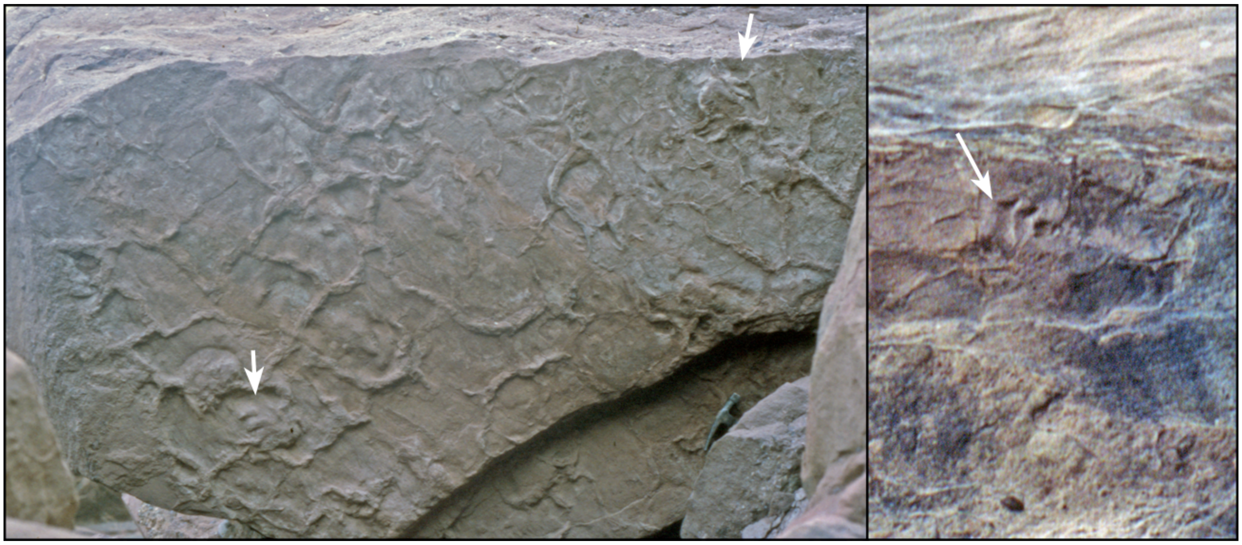


Figure 22: Natural casts of footprints and mud cracks from the Honeycomb Point Formation at Stop 1.1, with stratigraphic positions shown in Fig. 21: left, displaced block form "a"; right, undersurface of ledge *in situ* high in cliff from "b". Arrows point to tracks but there are more tracks than arrow in the left example.

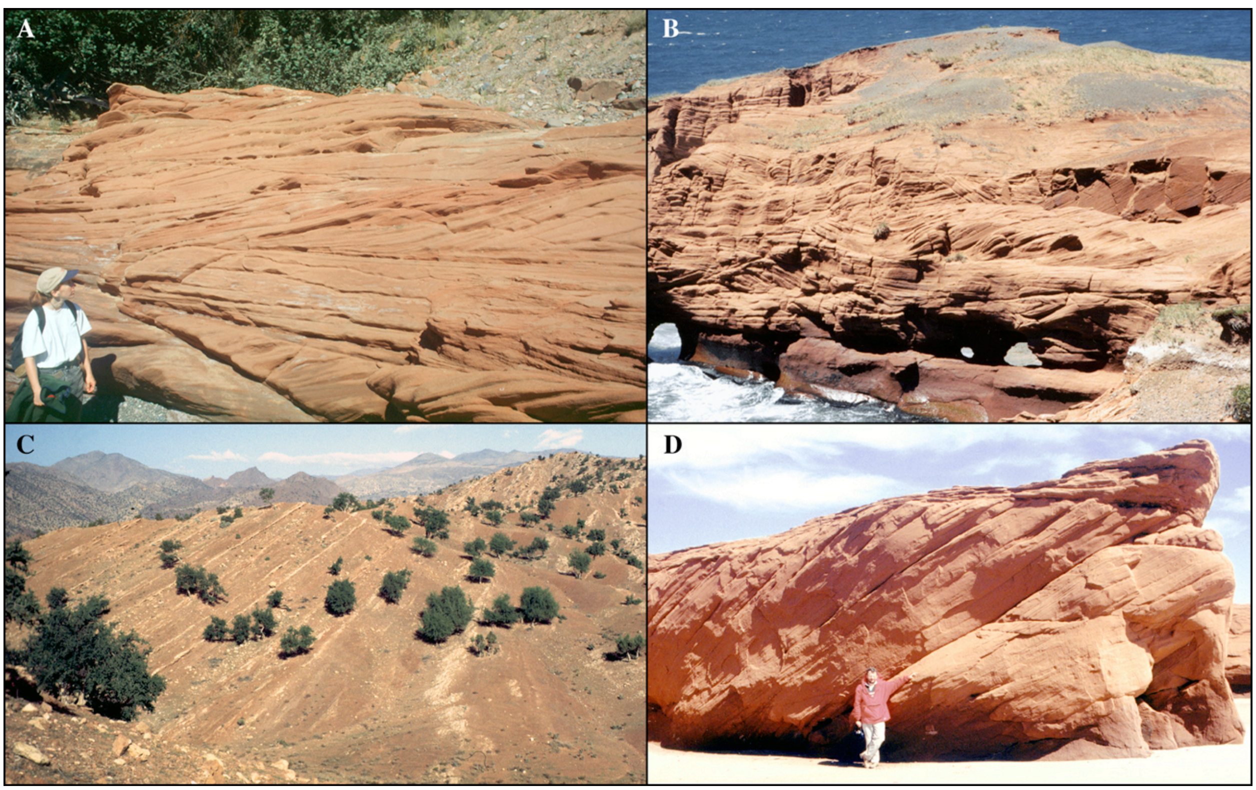


Figure 23: Comparison of facies of Honeycomb Point Formation eolian sandstone south of Quaco Head (A: Sarah Fowell for scale), Cap Aux Meules Formation eolian sandstone on the Isles de la Madeleine (B and D: cliff is about 7 m high in B and Roy Schlische for scale in D), and the non-eolian Ikakern Formation, east of Timesgadiwine, Morocco, with carbonate beds.

Stop 1.2: Macs Beach, St. Martins New Brunswick, Canada

45°21'31.02"N, 65°31'30.88"W

Main Points:

Juxtaposition of arid and relatively humid facies
Unconformity between TS I and TS II
Gray facies not visible on hanging wall
Locally oldest Triassic deposits in basin

Cliffs along the west side of St. Martins at "The Caves" at Macs Beach (Fig. 6) reveal about 70 m of brick red Honeycomb Point Formation, unconformably overlain by more than 100 m of the Quaco conglomerate (Powers, 1916). The Honeycomb Point Formation here, grouped into Nadon's (1981) McCumber Point Member lacks prominent eolian sands and instead has abundant gravely sandstone and conglomerate with generally angular clasts comprised predominately of local metabasalts and rhyolites. The stratigraphic relationship between these strata and those at Stop 1.1 is not obvious, however, because there are few facies types shared between units at Honeycomb Point and "the Caves", it seems likely that the latter occur at a higher stratigraphic level than the former within the formation as whole.

Footprints, first noticed by Donald Baird in the 1960s (pers comm.), occur as sole marks at the contact between mudcracked red mudstones and overlying red sandstones, and have been seen sporadically on the roofs of the "Caves". Like those that have been seen at Honeycomb point, the forms seen this far are all quadrupedal and look very different from typical Triassic forms.

Brick red Honeycomb Point strata are separated by a angular unconformity from the overlying gray and brown Quaco conglomerate, the dip of the Honeycomb Point beds being 2-3° greater than the overlying beds, consistent with tilting towards the border fault complex prior to the deposition of the Quaco. The Quaco conglomerate, as seen here is largely a clast-supported unit comprised mostly of very well rounded quartzite cobbles and, with finer grained intervals having occasional black comminuted plant fragments. Nothing like it is seen in the Honeycomb Point Formation anywhere. While the Honeycomb Point Formation was clearly deposited in an arid to very arid environment, with flashy streams, the Quaco was deposited by a river system with more sustained flow, suggesting a significantly more humid environment. The Quaco conglomerate fines upward grading upward into finer grained overlying strata termed the Echo Cove Formation by Nadon & Middleton (1984), varies between largely red and largely gray intervals (distinguished as members of the formation by Nadon & Middleton, 1984), which we will see at the next stop (Stop 1.3). The gray units contain organically preserved plant material, including pollen and spores indicating a Carnian age, and again suggesting a significantly more humid climate. While most authors have regarded both the Quaco and the Echo cove units as formations, we regard them as members of the Wolfville Formation (Wolfville Member of Powers 1916, raised to formation by Klein, 1962), broadly defined. Because the palynomorphs recovered thus far come from an interval more than 1 km above the base of the Quaco, the latter may be significantly older than Carnian, possibly Middle Triassic. In any case, the Quaco is the oldest Triassic deposit

Close examination of the abundant quartzite clasts from the Quaco Member shows two interesting features described clearly by Klein (1963). First, are rough pale craters, sometimes with radiating fractures on the cobble surfaces. When the cobbles are examined in situ, it is clear the craters occur at contact points with other clasts. Recently they have been reinterpreted as resulting from shock features resulting from the passage of a shock wave from an extraterrestrial impact, in this case the Manicouagan impact (Tanner, 2003). These are identical to features also interpreted as due to far field impact shock structures on quartzite cobbles from the Early Triassic "Buntsandstein" of Spain (Ernstson et al., 2001), and the Late Triassic Shinarump conglomerate of Arizona (Bilodeau, 2002). Conversely, Stel et al. (2002) and Chapman et al. (2004) (and of course Klein, 1963) conclude these features are due to pressure solution at point contacts. The pressure solution interpretation is supported by the very widespread nature of the features in other units. Three additional examples we have observed are: 1, the Late Cretaceous/Paleocene North Horn Formation conglomerates in Spanish Fork Canyon, Utah; 2, Early Jurassic post-rift conglomerates near Bigoudine, Taroudant, Morocco; and 3, Late Triassic Passaic Formation, at Mt. Ivy, New York. In fact, we have observed these types of craters on every quartzite-bearing clast-supported conglomerates we have looked at. Second, there are crescent-shaped fractures on the surfaces of many of the same Quaco cobbles. These however are not a contacts between cobbles and have no relation to other cobbles. These are percussion marks, described in detail by Klein (1963) and seen today in environments where cobbles are rapidly transported.

The Quaco Member is most likely apparently a local unit, deposited by a large river entering the basin early in the depositional history of TS II. Its importance (and that of the overlying Echo Cove Formation) is in its contrast to the underlying Honeycomb Point Formation plausibly caused by a significant change in plate position between the Late Permian and the Middle to Late Triassic. Based solely on climate sensitive aspects of the sedimentary facies and assuming a constant zonal climate and Permian placement the transition seen between the Honecomb Point and the Quaco represent a southward translation of the North American plate and

Return to vehicles. (About 15 minutes to Stop 1.3)

Return to Big Salmon River Road and turn right (northeast) toward Little Beach Rd	2.8 km
Continue on Little Beach Road to Fundy Trail Parkway Entrance	4.4 km
Proceed along winding road northeast to Parking area P2	1.7 km
Park and walk northeast to trail head to Melvin Beach	0.2 km
Follow Trail to Melvin Beach	0.5 km

Melvin Beach: 45° 23.794'N, 65° 27.177'W

Stop 1.3: Melvin Beach, Fundy Trail Park, New Brunswick, Canada

45° 23.794'N, 65° 27.177'W

Main Points:

- Finer clastic sequence continuing the more humid climatic regime of Quaco
- Preserved thickness greater than entire TS II on outcropping hanging wall
- Gray strata abundant, representing a facies typical of TS II along western edge of basin
- No sign of lacustrine strata

Outcrops to the east and west of Berry Beach consist of cliffs of the Echo Cove Member of the Wolfville Formation. The Echo Cove Member overlies the Quaco Member and consists mostly of gray, tan, and red sandstone and gravelly sandstone with less abundant red and gray mudstone. The gray units contain organically preserved plant material and pollen and spores. About 1 km of strata is present in these outcrops and those to the west to Echo Cove itself.

The Echo Cove Member was deposited by a fluvial system consisting of smaller streams than the underlying Quaco Member (Nadon, 1982; Nadon & Middleton, 1984). There is no sign of either deltaic or lacustrine units, although it is possible that these strata could be lateral equivalents of lacustrine strata deeper in the basin (cf., Wade et al., 1996). This hypothetical relationship has not been tested, however, because strata of TS II were not encountered by the two exploratory wells in the basin.

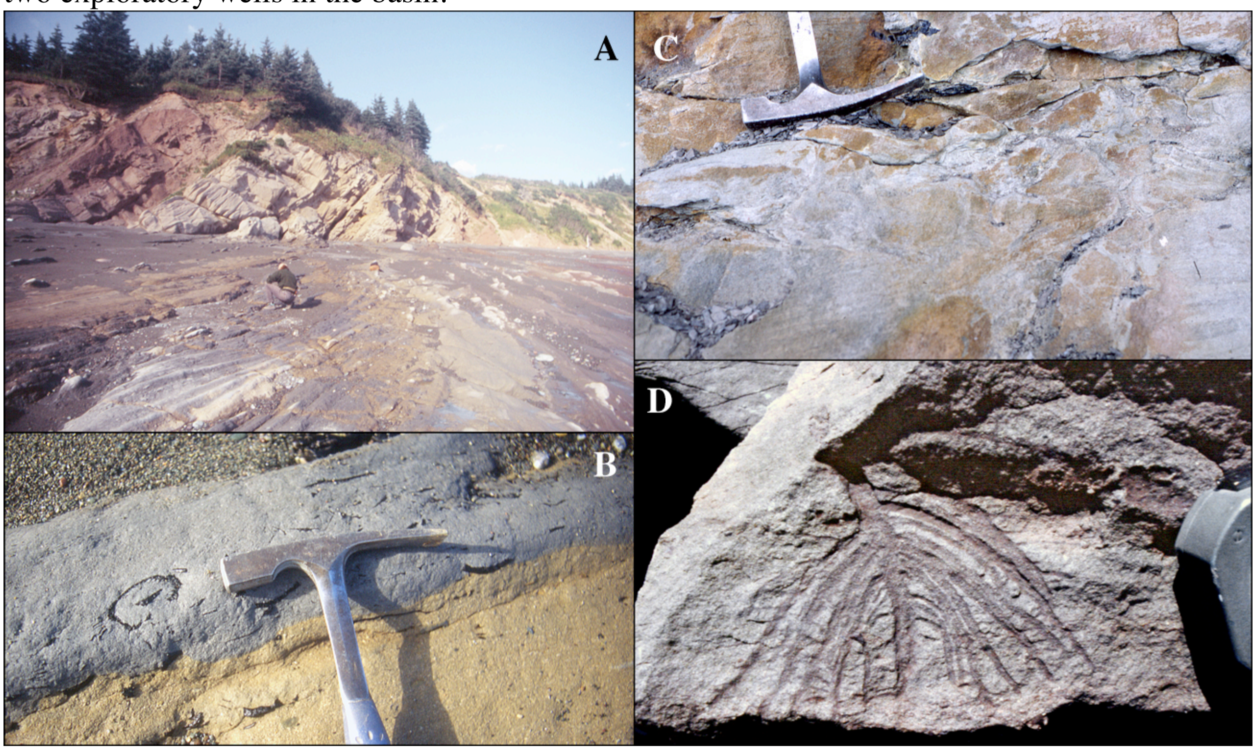


Figure 24: Plant-bearing strata of TS II (Echo Cove Mb., Wolfville Fm.) at Martin Head (A.B), and in Stop 1.3, Melvin Beach: A, Gray and brown strata at Martin Head (Sarah Fowell for scale); B, ?*Neocalimites* stem and wood fragments in gray sandstone at Martin Head; C, wood in gray sandstone at Fownes Head, Melvin Beach; D, root traces of unknown plant in red sandstone, east side of Melvin Beach.

On the west side of Berry Beach is Fownes Head. Strata in the cliff face are largely gray and have produced pollen and spores as well as comminuted plant fragments and wood (Fig. 24). This is the type section of what Nadon (1982) and Nadon & Middleton (1984) term the Fownes Head Member, which overlays a largely red unit gradational into the underlying Quaco they term the Berry Beach Member. On the east are outcrops of red clastic rocks that Nadon (1982) and Nadon & Middleton (1984) term the Melvin Beach Member (Fig. 24), which they regard as overlying their Fownes Head Member. However, superposition of their type Melvin Beach Member over the Fownes Head Member cannot be demonstrated here, and because conglomerates resembling the Quaco occur below the Melvin Beach unit to the east, we argue the Melvin Beach unit is equivalent to the Berry Beach unit. In any case, we regard as these units as informal divisions of the Echo Cove Member, because they cannot be mapped outside of this area. If this interpretation is correct than the Echo Cove Member becomes predominately gray upward, without a return to well-developed red beds within these outcrops.

The Fownes Head beds have produced pollen and spores indicating a Carnian age (Traverse, 1983; Cornet in Nadon & Middleton, 1984; Wade et al. 1996), largely based on the presence of the pollen taxon *Patinasporites* which is virtually unknown in strata older than Carnian in Europe, with notable exceptions (e.g., Courtinat & Rio, 2005). This age is consistent with that based on vertebrates (Olsen et al. 1989).

Table 1: Pollen and spore taxa from Martin Head and Fownes Head. Samples provided by D. H. Magnusson and D.E. Brown and processed by R.P. Stapleton, 1987 (courtesy of D.E. Brown).

Age diagnostic taxa

Carnisporites cf. *C. mesozoicus* (Klaus) Mädler 1964
Cyclotriletes oligogranifer Mädler, 1964
Leschikisporis aduncus (Leschik) Potonié, 1958
Lundbladispora sp.
Tigrisporites sp.
Aratrisporites virgatus (Leschik) Mädler, 1964
Enzonalasporites leschiki Mädler, 1964
E. tenuis Leschik, 1955
Lueckisporites sp.
Minutosaccus potoniei Mädler, 1964
M. schizeatus Mädler, 1964
Ovalipollis sp.
Patinasporites tardis Leschik, 1955
Taeniaesporites sp.

Baculatisporites. sp.
Klukisporites sp.
Lycopodiumsporites sp.
Punctatisporites sp.
Retusotriletes sp.
Verrucosisporites sp.
Alisporites sp.
Cedripites so.
Klauslpollenites sp.
Pinuspollenites sp.
Platysaccus sp.
Vitreisporites sp.
Froelichsporites traversei Litwin et al. 1993 [1]

[1] Martin Head only (added by PEO)

Addition forms

A major difference between TS II on this shore, and the outcrops along the strike slip and hinge margin of the basin are the relative abundance of plant-bearing gray strata on the former in TS II with two other outliers exhibiting facies that are variations on the Echo Cove Member. Outcrops at Martin Head (Fig. 24) consist of gray, purplish, and red sandstones and less common mudstones. This site, access to which is difficult, has produced a variety of macroscopic plants and in situ plant stems can be seen in some of the yellow-weathering, gray sandstones (Fig. 24). Martin Head has produced a playnoflora that is consistent with that from the Echo Cove Formation indicating a Carnian age (Table 1). This assemblage is also consistent with palynoflorules from the Oukaimeden sandstones formation of the HABC indicating a Carnian age (Cousminer & Manspeizer, 1976; Biron & Courtinat, 1982). Strata at Martin Head are fluvial in origin, lacking eolian beds, and have even higher proportion of gray beds than the Echo Cove Formation consistent with a considerably more humid environment. Fluvial strata at Waterside are predominately red and brown, but plant stems and tree branches without organic preservation are abundant and eolian strata are lacking consistent with the style of sedimentation Echo Cove Formation. We therefore group these other two outliers with the Echo Cove Member of the Wolfville Formation and conclude that the Echo Cove style of sedimentation is probably representative of much of the Wolfville close to the northwestern edge of the basin.

In general the Echo Cove Member and its equivalents in New Brunswick resemble the sandstone sequences in TS II of the Argana basin (Irohalen Member of the Timesgadiwine Formation) and the Oukaimeden sandstones formation of the HABC. It is quite possible, furthermore, that some of the Echo Cove strata might grade laterally into lacustrine strata deeper into the basin as seen in time-equivalent strata in the Argana Basin (Hofmann et al., 2000). What these lacustrine strata might look like, and what their source rock quality might be if indeed they exist is completely conjectural, given the fact they have never been sampled in the subsurface.

Thermal modeling of apatite fisson track data from sandstones from Stops 1.1, 1.2, and 1.3 and Martin Head (Fig. 25) suggests a late stage of heating during the Paleogene (Grist & Zentelli, 2003). This is potentially very important for the Bay of Fundy hydrocarbon system because while peak hydrocarbon generation and mobilization from pre-rift sources was probably during the early Jurassic the Paleogene reheating could have remobilized hydrocarbons into traps that have since been stable. Thermal Alteration Indices (TAI) of the Fownes Head and Martin Head are 2 1/2 and 3 (early mature) according to Stapleton so that the section should still be in the oil window (Table 1).

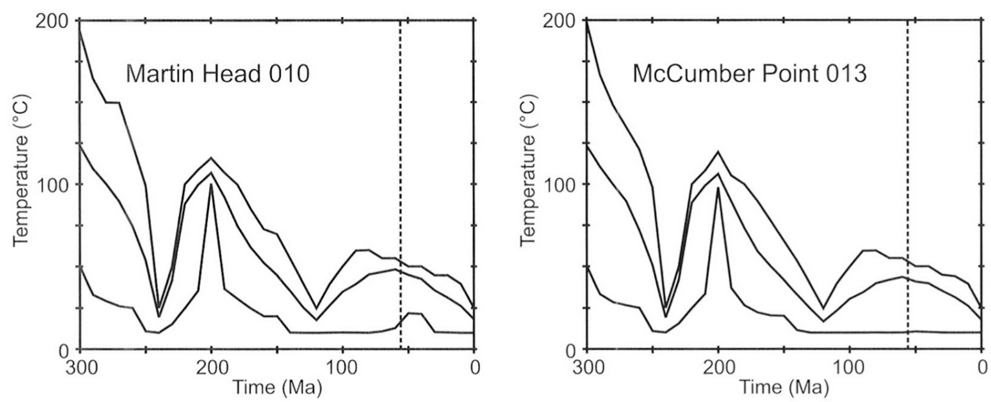


Figure 25: One heat pulse models for apatite fission track data from Stop 2.1 (right) and Martin Head (left), from Grist and Zentelli (2003).

Return to vehicles at Parking area P2 (about 0.7 km). (About 1:00 hr to Stop 1.4)

Retrace drive back to parkway entrance	1.7 km
Continue on Little Beach Road to Big Salmon River Road	4.4 km
Follow Big Salmon River Road towards Main Street, St. Martins	3.7 km
Continue on Main Street	2.3 km
Turn right at 111	17.5 km
Entering Moncton basin	
Continue on 111	8.4 km
Turn left at 865	17.5 km
Turn right to merge onto 1 E	1.5 km
Pull over onto shoulder: Stop 1.4 just past 177 km marker	

Stop 1.4: Albert Formation: Carboniferous age prerift source rock for McCully gas and Stony Creek oil fields, Moncton basin, New Brunswick, Canada

45° 38.591'N, 65° 40.624'W

Main Points:

- Carboniferous age lacustrine sequence
- Potentially underlying parts of Fundy basin
- Deep to shallow water lacustrine facies
- Albertite veins

The Moncton basin is a transpressional basin formed during the docking of African and North American plates during the Carboniferous, one of several that formed in the Maritimes of Canada that flank and underlay parts of the early Mesozoic Fundy rift basin. The Albert formation is a Mississippian age lacustrine sequence within the Moncton basin, one long know for its relatively low thermal maturity organic rich rocks that in addition to being source beds for hydrocarbons, also have produced rich fossil assemblages including the palaeonisciform fish *Rhadinichthys, Elonichthys*, and *Canobius* from the richest of the oil shales, and disarticulated remains of palaeonisciforms and the crossopterygian *Latvius* sp. from less well laminated intervals. Conchostracans and ostracodes are present and plant remains are abundant, including fossil forests described by Falcon-Lang (2004) that are 2 km to the west along this highway. Keighley & Brown (2005) point out that hydrogen/oxygen indices of the fish-bearing oil shales identify the kerogen to be a low maturity type 1, with unusually low oxygen content. They note that this is typical of a primarily lipid rich, algal source as would be expected in a lacustrine basin.

Exposures at this stop (Fig. 26) have been described by Keighley & Brown (2005: Stop 2.5) and consists mostly of gay to dark gray laminated mudstone and fine-grained, cross-laminated, irregularly laminated, or massive sandstone interpreted as lacustrine mud and sand-flat and perennial lacustrine. There are significant, cross stratified, sharp based, light grey, highly feldspathic pebbly sandstone which are plausibly fluvial in origin at the top and bottom of this section. Small veins of solid hydrocarbon (presumably albertite) are present within the laminated mudstone around 10 metres (Fig. 26). Rocks in the vicinity of this stop have undergone significant deformation as is typical for the finer-grained deposits within the Albert Formation.

The lacustrine shales of Albert Formation have apparently sourced sandstone reservoirs within the same formation producing commercial accumulations in two areas. First, is the McCully gas field, which we will pass through after leaving this stop. The field is described by Keighley & Brown (2005) from which the following is paraphrased. The field was discovered in 2000 by Corridor Resources Inc./Potash Corporation of Saskatchewan in the McCully #1 well east of Sussex. The discovery well produced gas from a 22 m sandstone within the lower Albert Formation and flowed at 2.5 mcf per day. The field covers over 7000 acres and is in a large anticlinal structure with simple four way closure. To date, drilling has been concentrated on the northwest limb. Gas pressure is consistently over 500 psi above the hydrostatic pressure at this depth, indicating a >500 m gas column. However, sandstone porosities (~ 8%) and permeabilities are low. The field is now in production, originally supplying power to the

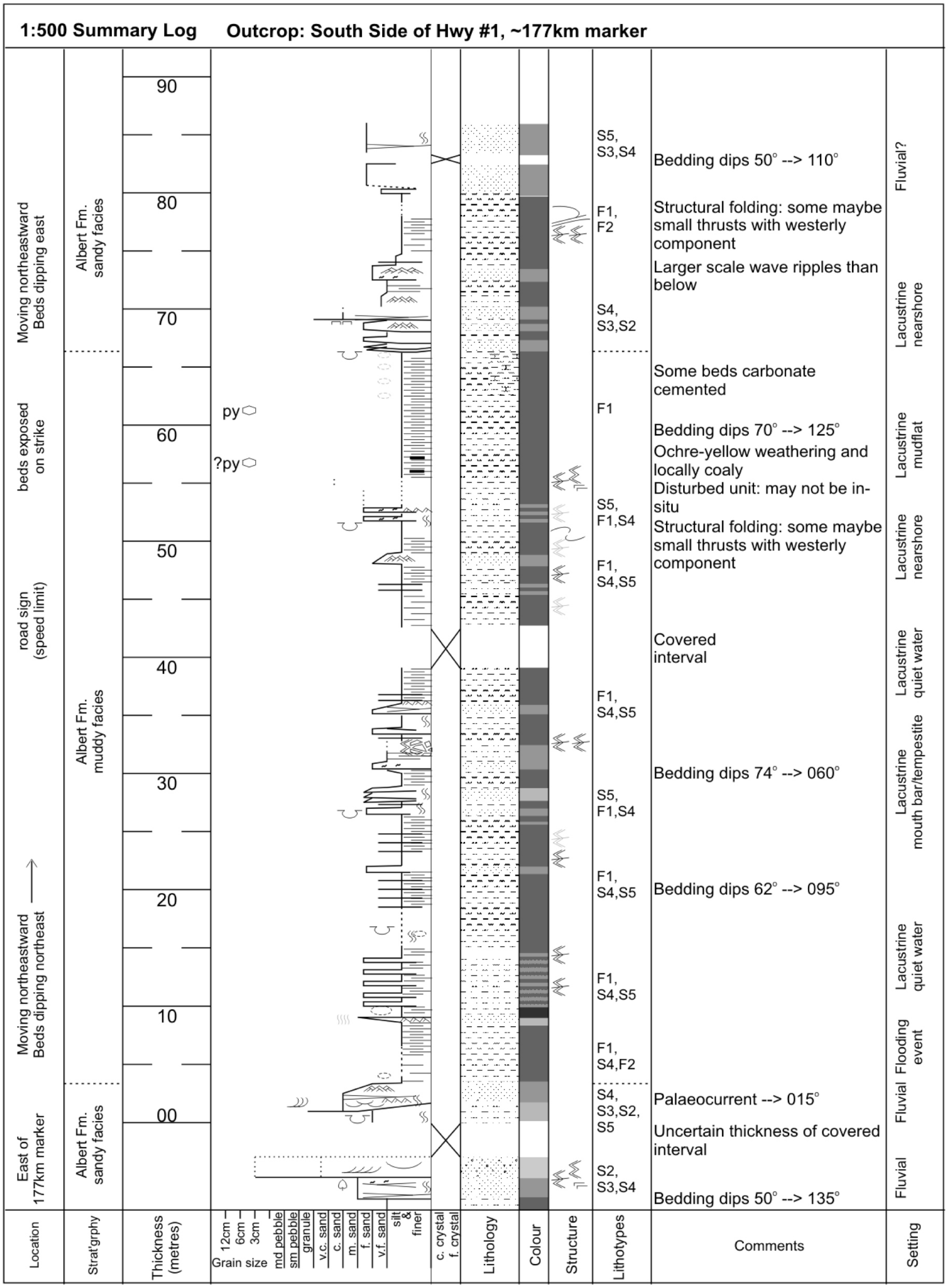


Figure 26: Sedimentary log of the section exposed adjacent to the Highway 1, 177 km marker, Stop 1.4 reduced from the original 1:500 scale (Reproduced from Keighley & Brown (2005).

adjacent potash mill, but recently a 35 km lateral feeder has connected the field since June 27, 2007 to the Maritimes and Northeast Pipeline. The proven and probable reserves, in just the northwest flank of the field, has been estimated at 119 bcf (Keighley & Brown, 2005)..

The second area is the old Stoney Creek oil and gas field south of Moncton, the description of which we have also paraphrased from Keighley & Brown (2005). Small quanities of oil were produced from wells near Dover around 1859 to 1905. In July 1909, Maritime Oilfields Ltd. of London, England, discovered the Stoney Creek field along strike with the Dover wells on the other side of the Petitcodiac River. In total, about 28.7 bcf of gas (~600,000 mcf per year from 1914 to 1947) and over 800 thousand barrels of oil were produced from the field and piped to nearby Hillsborough and Moncton from 1912 until the 1970s until closure in 1991. Total oil in place is calculated at ~2.1 x 106 cubic metres with <5% primary recovery. The Albert Formation reservoirs are in sandstone lenses up to 30 m thickat depths ranging mostly from 600m to 900m interbedded and overlain by lacustrine mudstone and oil shale in a stratigraphic trap consisting pinches out up-dip into mudstones. More recently, a gas discovery (the Downey prospect) was made 5km south of the Stoney Creek field in 1998 by MariCo Oil and Gas Corp., exploration being sparked by the construction of the Maritimes and Northeast Pipeline to transport offshore (Sable Island) natural gas to the Boston area.

The Albert Formation is one potential source rock that could be buried below the Fundy basin. The fact that the basin may have experienced elevated temperatures in the Paleogene (Stop 1.3) makes it quite possible that a pulse of hydrocarbon generation and mobilization may have occurred that recently without subsequent disruptive tectonic events. The most likely area of the Fundy basin to have Moncton basin strata, including the Albert Formation, in the subsurface is in the Chignetco subbasin, and area that has never been explored.

Return to vehicles. (About 2:42 hr to Parrsboro, Nova Scotia overnight location).

Proceed northwest on Route 1 east.	28.0 km
From this area you can see the Potash Corporation of Saskatchewan plant and the McCully gas field at Penobsquis, NB, continue on 1 E	37.1 km
Merge onto 2 E Entering Nova Scotia	89.7 km
Continue on 104 E go 8.2 km	
Keep right at the fork, follow signs for 2 S/Glooscop Trail/Springhill/Amherst/Maccan and merge onto 2/Glooscap Trail	1.5 km
Sifto Salt Plant on right, continue on 2 S	1.0 km
Turn right at 302/Fundy Shore Scenic Dr/Glooscap Trail	
Continue to follow 302	25.2 km
Continue straight merging onto Hwy 2	25.2 km
Arrive center of Parsborro, overnight stop	

Field Day 2: Strike Slip Margin of the Fundy Basin

Outcrops on the north side of the Minas physiographic basin (Fig. 27) cover the same basic sequence we saw on our traverse down section on Field Day 4. However, the section abuts the Minas fault zone, which was apparently active during sedimentation. Our stops are organized to take advantage of the tides. Stops 2.1 (Partridge Island) and 2.4 (Fundy Geological Museum) are independent of the tides. Stop 2.3 (Five Islands) is completely tide-dependent and while Stop 2.2 (Wasson Bluff) is less so. Stop 4.4 (Fundy Geological Museum) is not influenced by the tide. Both stops involve long-long walks. Fieldtrip participants should have footgear that can get wet and should be prepared for chilly weather in the event of cloudy, rainy, or windy conditions.

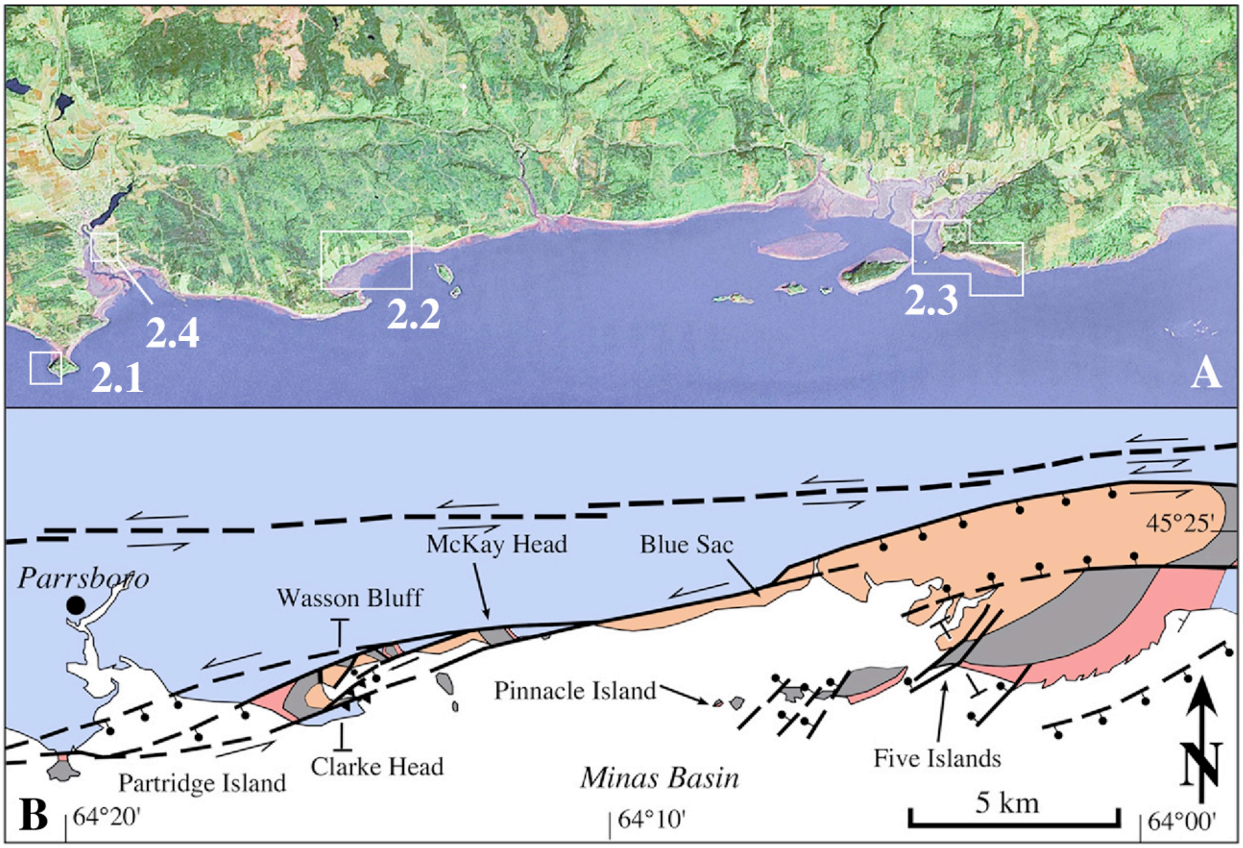


Figure 27: Index map for stops along the north shore of the Minas physiographic basin. A, Satellite photograph of field area with field stops shown as white boxes 2.1 – 2.4 (2005 Globexplorer). B, Detailed geological map of field area showing areas discussed in text and positions of cross sections shown in Figure 37 (Wasson Bluff to Clarke Head – Stop 2.2) and Figure 43 (Five Islands – Stop 2.3) and (adapted from Withjack et al., 1995). Position of map in Figure 37 is white box next to 2.2. Key to rock units in Figure 18.

Begin Parrsboro center, continue southeast on Main Street	0.2 km
Continue southeast on Lower Main Street	0.4 km
Continue on Whitehall Rd	3.2 km
Turn sharp left onto unnamed road (loop from Ottawa House)	130 m
Turn right onto road to beach	0.7 km
Park and proceed west	160 m
Turn to the southeast and walk along cliff edge	90 m
Arrive Stop 2.1	

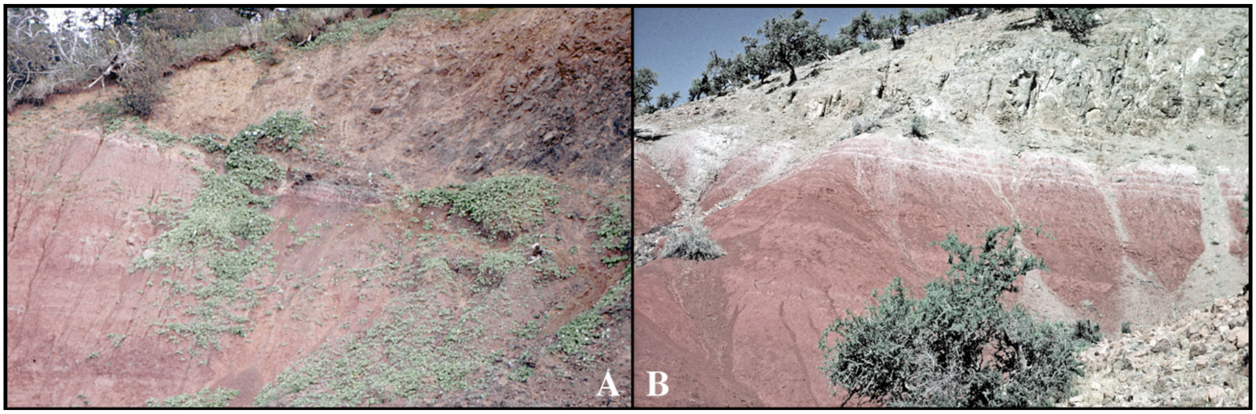


Figure 28. A, Triassic-Jurassic section at Partridge Island, view looking east towards cliff face with North Mountain Basalt above, the white-weathering bands of grey and black mudstone interbedded with red mudstone of the Partridge Island member of the Blomidon Formation, below and the underlying Whitewater Member of the Blomidon Formation. B, Triassic-Jurassic section near Argana, Morocco, view looking to the west, with the Argana Basalt overlying the Bigoudine Formation.

Stop 2.1: Partridge Island: Triassic-Jurassic Boundary

45° 22.109'N, 64° 20.238'W

Main Points:

Excellent and accessible outcrop of the Triassic-Jurassic boundary
Pollen and spore extinction level
Iridium anomalies
Gabbroids in basalt
Comparisons to Morocco

Park along the gravel tombolo connecting Partridge Island with the mainland and proceed to the west face of the "island". On the west side is an outcrop of the uppermost Blomidon Formation and lower half of the North Mountain Basalt (Fig. 28). It is a relatively easy climb up to the contact, which is highly recommended. It has had a tendency to get a bit overgrown in recent years so some digging may be required.

. The lower 15 meters of section at Partridge Island consists primarily of red mudstone. Sand patch cycles are not at all obvious. The upper 1.5 m of the section below the North Mountain Basalt consists of the red, purple, grey, and black thin beds that comprise the Partridge Island member (Figs, 28, 29). When not metamorphosed, this member is highly palyniferous, as it is here at the type locality of the informal member.

The Partridge Island section was first described by Fowell & Traverse (1995) who showed that the last appearances of a number of typical Triassic taxa occur in the uppermost 20 cm of the Blomidon. According to Fowell & Traverse (1995), the well-preserved palynofloras are dominated by the genus *Classopollis* (née *Corollina)*. The less common elements in the uppermost assemblages include species that are also present in earliest Jurassic assemblages from the Hartford basin. Palynomorph assemblages 30 cm down section from the basalt contain rare specimens of the Late Triassic index species *Patinasporites densus* and a series of monosulcate grains shared with late Rhaetian assemblages from the Newark basin (Figs. 30, 31). Their analysis indicates that the Triassic-Jurassic boundary is within the uppermost few tens of centimetres of the Partridge Island member

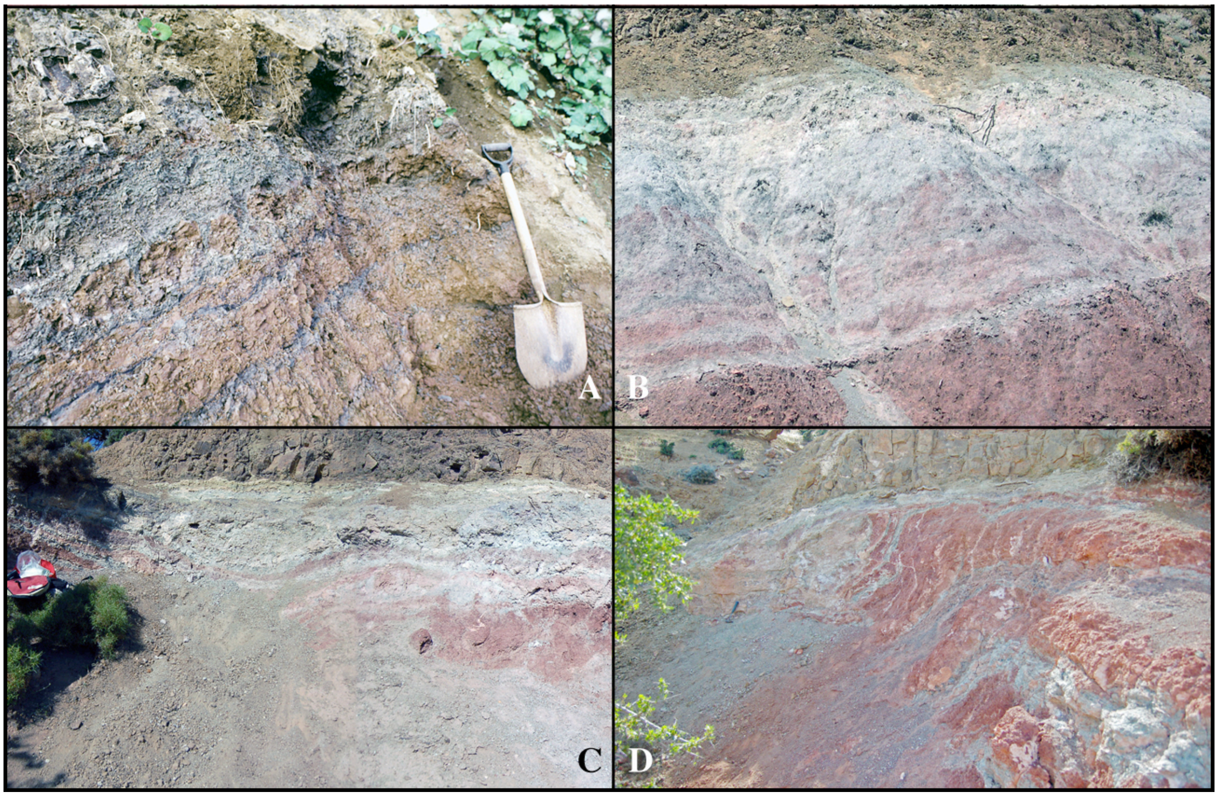


Figure 29: Detail of the Triassic-Jurassic boundary interval in Nova Scotia and Morocco: A, Partridge Island, Stop 2.1, showing organic-rich bands (1-5% TOC) alternating with red, some with large grey-filled cracks (base of the North Mountain Basalt is at arrow: shovel is about 1 m long); B-D, sections at Igounan, Argana basin, Morocco: B, Relatively thick boundary interval, total grey bed thickness about 2.8 m thick with beds being even and laterally continuous; C and D, section to south of A showing salt-dissolution bowls showing growth.

A problem with this section, and indeed all low accumulation rate boundary sections in the Fundy basin and in Morocco (c.f., Olsen et al., 2003a,b; Marzoli et al., 2004) is that there are often subtle low angle thrust and normal faults that distort the section in ways often very difficult to discern. Coupled with depositional lateral variations, including mud cracks (see Fig. 29), repeated attempts at measuring and sampling the same section are uncomfortably discrepant. Such repeated measurements provide an object lesson in stratigraphic precision, and strongly suggest analytical measurements should be done on the same samples collected in one sampling effort to assure registry among the data sets. One such section (by P.E.O.) was reanalyzed

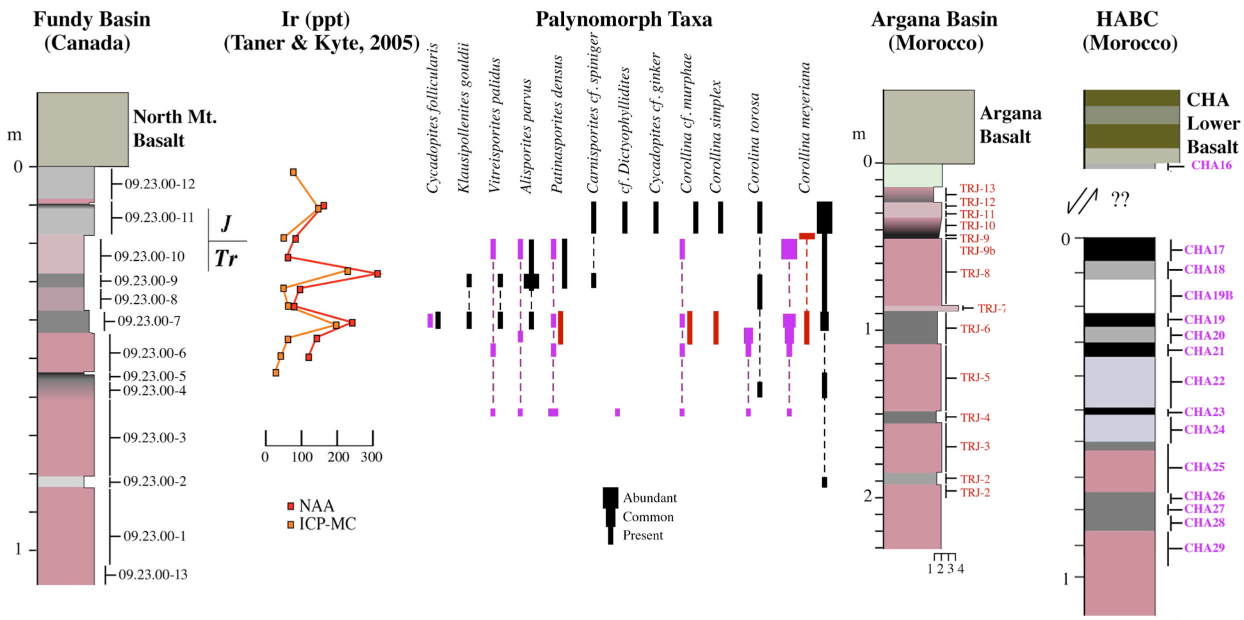


Figure 30: Comparison of Triassic-Jurassic boundary sections and palynological data in the Fundy basin of Nova Scotia and the Argana and HABC basins in Morocco. Palynomorph processing and identification by S. Fowell and B. Cornet.

by Fowell (in Olsen et al., 2005a; Whiteside et al., 2007) (Fig. 30) revealing the same basic pattern reported by Fowell & Traverse (1995), but with slightly different unit thicknesses. This pollen transition is consistent with what is seen in the Triassic-Jurassic boundary interval Newark basin (Fig. 31).

There are several ways to estimate the accumulation rate at this section. One way is to use the magnetic reversal stratigraphy in the GAV-77-3 core from Margaretsville, Nova Scotia (Kent & Olsen, 2000). The Partridge Island Member was represented in this core, but was decimated by sampling, however it is still relatively well preserved in the adjacent core AV-C-1-4 (Sladen [Quebec] Ltd. cored in1966 at Margarettsvile) (Fig. 32). Correlating to the Newark basin astronomically calibrated GPTS) (Fig. 33), it is clear that the accumulation rate is slowing through the history of the Blomidon Formation. Fitting an exponential curve to the all of the data, the best fit (Fig. 32) predicts an accumulation rate of 1.65 m / 100 KY for the top of the Blomidon Formation. If we use only the first possible magnetic polarity tie point (correlative of middle of E22r) and the base of the basalt, we get 0.76 m /100 ky. In the Newark basin the Triassic-Jurassic boundary is estimated to occur between 1 and 2 Van Houten cycles (20 to 40 ky) below the base of the North Mountain Basalt. Based on these two estimates, the Triassic-Jurassic boundary should be between 33-66 cm to 15-30 cm below the boundary. The observed position of the boundary between 20 and 30 cm below the basalt is certainly in line with these estimates.

Another way to estimate the accumulation rate is to correlate the cyclostratigraphy of Partridge Island member and the Newark basin independent of the magnetostratigraphy. Assuming that the Partridge Island member correlates overall to the Exeter Member of the Newark basin section, there is a fairly good match between the individual Van Houten cycles (Fig. 15). Thus estimated, the accumulation rate is between 20 to 35 cm per 20 ky cycle. An estimate of 20 cm / 20 ky was derived this way by Olsen et al. (2002b).

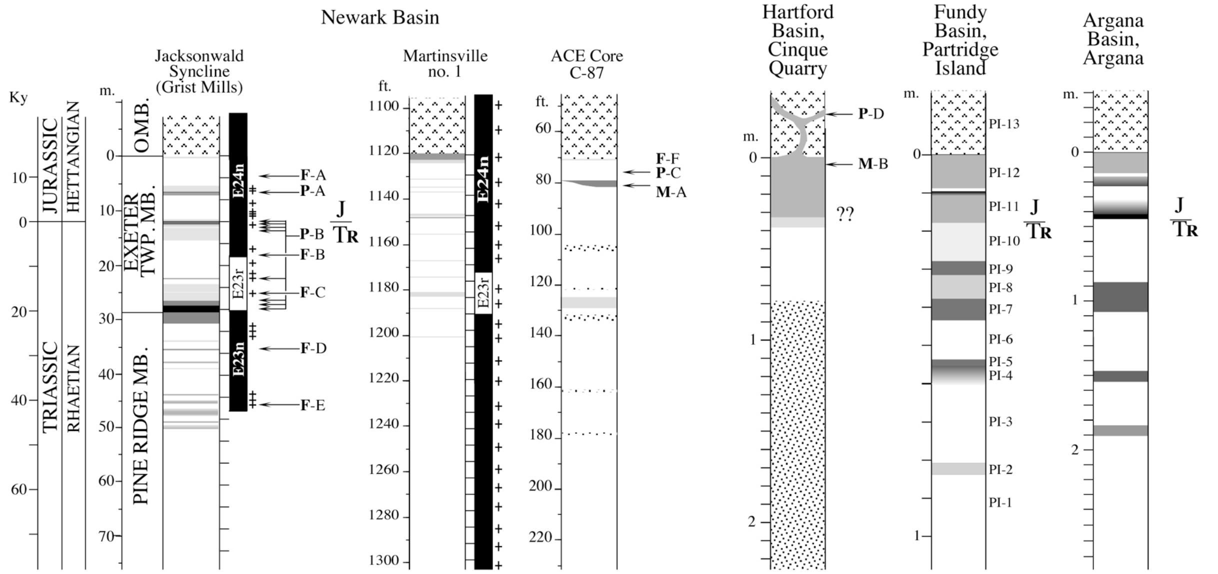


Figure 31: Newark, Hartford, Fundy, and Argana basin Tr-J sections (from Olsen et al., 2002b). Abbreviations: **F**-A, brontozooids only; **F**-B, *Batrachopus deweyii*, *Rhynchosauroides*, and small brontozooids; **F**-C, last appearance new taxon B and *Brachychirotherium* and cf. *Apatopus*, *Rhynchosauroides,* and small brontozooids; **F**-D, *Brachychirotherium*, new taxon B, *Batrachopus*, and small brontozooids; **F**-E, abundant *Brachychirotherium*, new taxon B, *Batrachopus*, and small to medium sized brontozooids; **F**-F, small to large brontozooids, including lowest occurrence of *Eubrontes giganteus*, also *Batrachopus deweyii*, and *Rhynchosauroides*; **P**-A, lowest definitive Jurassic-type palynomorphs assemblage; **P**-B, palynomorph assemblages of Triassic aspect (lower) or dominated by spores (upper); **P**-C, palynomorph assemblage with *Classopollis* only; **P**-D, palynomorph assemblage with *Classopollis* only in matrix between basalt pillow; **M**-A, macrofossil plant assemblage dominated by *Brachyphyllum* and *Clathropteris*; **M**-B, macrofossil plant assemblage dominated by *Brachyphyllum*.

Thus, independent, non-biostratigraphic, estimates of accumulation rate suggest that the palynological extinction level and boundary cyclostratigraphies in the Newark and Fundy basins are correlative. This is important because there are a variety of other possible proxies of environmental change that are shared between the two basin sections across the boundary that will be discussed at the outcrop.

Repeated sampling has resulted in an emerging picture of correlated environmental changes. First, the palynological transition in the Fundy basin is marked by an Ir anomaly of up to about 300 ppt (Tanner & Kyte, 2004; pers comm.). An Ir peak of the same magnitude is also seen in the Newark basin section exactly at the palynological transition (Olsen et al., 2003a; Whiteside et al., 2003). There are also new exciting carbon isotopic data that will discussed on the outcrop. Given the consilience of the accumulation rate and cyclostratigraphic information, and the parallel palynological, Ir and other chemostratigraphic data, a regional if not a global set of processes are plausibly responsible. The details of these observations will be presented with data in the field.

Another boundary section showing different but related phenomena is at Central Clarence, Nova Scotia, within the Partridge Island Member (Fig. 34). This is the only macrofossil plant locality in the Blomidon Formation, and the identifiable floral assemblage consists entirely of the fern *Cladophlebis* (Baird in Carroll, 1972). A potential Ir anomaly is associated with this section (Orth in Mossman, 1998). Although the contact with the North Mountain Basalt is covered, the fern horizon and Ir peak and may represent another example of a “fern spike” similar to that seen in the Newark basin (Olsen et al., 2002a,b).

The Central Clarence section is much coarser than at the Partridge Island section. This is part of a general trend within the Partridge Island member along strike of the basin, from Partridge Island to Digby Gut. Facies and thicknesses within the Partridge Island member are basically similar from Partridge Island to Cape Blomidon to the GAV-77-3 core. But from there to the west, the member coarsens and thickens dramatically, so that at Digby Gut, it is a grey conglomerate. Since no conglomerate is present in the position of the Partridge Island member within the Cape Spencer or Chinampas wells or at Grand Manan Island, we presume that this indicates a source of coarse sediment to the south from the hinge margin, consistent with accelerated tilting just prior to the extrusion of the North Mountain Basalt.
Olsen et al. (1987, 1990, 2002a,b) have argued that the available data are consistent with the impact of a bolide, in a scenario similar to that of the K-T boundary. Indeed the parallels are impressive: extinction level (Cornet, 1977; Fowell & Olsen, 1993; Fowell & Traverse, 1995), Ir anomaly (Olsen, 2002a; Tanner & Kyte, 2004), fern spike (Fowell et al, 1994), negative $\delta^{13}C$ excursion (Hesselbo et al., 2002; Ward et al., 2004), and CO_2 anomaly inferred from stomatal density changes (McElwain et al., 1999; Beerling & Berner, 2004). Shocked quartz has been reported as well (Bice et al., 1992), although not corroborated by the others. However, the temporally close CAMP basalt flows, analogous with the Deccan Traps of the Cretaceous-Tertiary boundary and the Siberian Traps of the Permo-Triassic boundaries, suggest another hypothesis.

Figure 32: Portion of core AV-C-1-4 (730-760 ft) spanning contact between the North Mountain Basalt and the complete local expression of the Partridge Island Member (747.0-756.6 ft.), and the uppermost Whitewater Member of the Blomidon Formation. The arrow marks the basalt-Blomidon contact. Core is youngest in upper right.

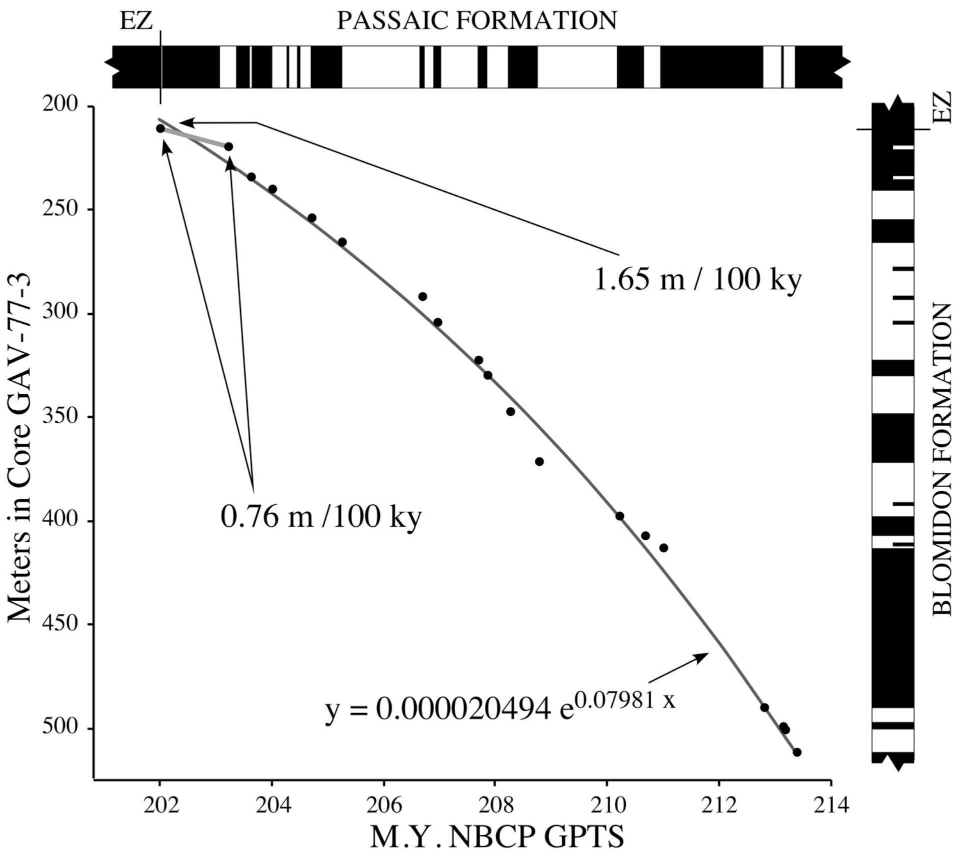


Figure 33. Correlation between Newark Basin Coring Project Astronomically Calibrated Geomagnetic Polarity Time Scale (Kent & Olsen, 1999) with the GAV-77-3 core of the Blomidon Formation and North Mountain Basalt (Kent & Olsen, 2000), and calibration of accumulation rates in the Fundy basin boundary section.

In contrast with the bolide hypothesis, the currently popular argument is that the CAMP eruptions, a portion of which is the North Mountain Basalt, somehow triggered a catastrophic change in climate, perhaps to a super-greenhouse (McElwain et al, 1999), with the release of methane from clathrates (Hesselbo et al., 2002; Beerling & Berner, 2004), or fatally cold climes via sulfate aerosols (McHone & Puffer, 2003). In any case the correlation between flood basalts and extinctions appears compelling (c.f., Rampino & Stothers, 1988; Rampino & Haggerty, 1996; most Hames et al., 2003), and the Ir anomaly is easily matched by basaltic ashes (Schmitz and Asaro, 1996), but at least for the Triassic-Jurassic boundary, there is no compelling evidence of any basalt extrusion prior to or at the time of the extinctions of the associated negative $\delta^{13}C$ anomaly. All of the places in which the CAMP flows can be seen in the same section as the extinction level or earliest Jurassic strata, CAMP flows occur above – never below. However, recently Marzoli et al. (2004) and Knight et al. (2004) have argued that a significant fraction of the basalts in Morocco predate the boundary. The Fundy basin is particularly germane to these arguments because most of the Moroccan basins have stratigraphies and basalt geochemistries remarkably similar to the Fundy basin.

One set of data used by Marzoli et al. (2004) to support the existence of pre-boundary basalts is geochemical data from the basalts. Marzoli et al. (2004) and Knight et al. (2004) argue that that the lower basalt unit of the Central High Atlas Mountains (CHA) of Morocco has higher

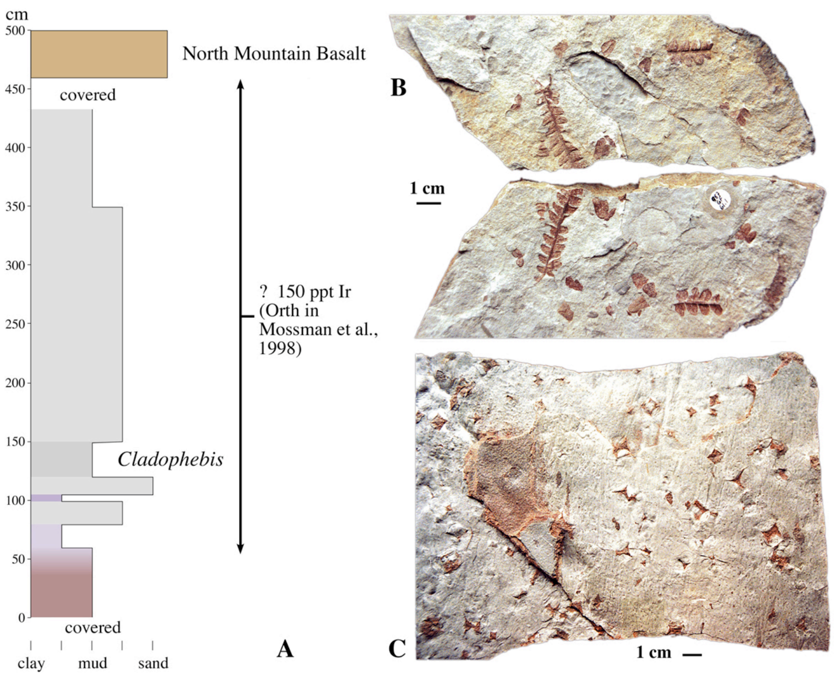


Figure 34: A, Triassic-Jurassic boundary section exposed at Central Clarence, Nova Scotia, showing the position of the *Cladophlebis*-bearing unit. B, Part and counterpart of a specimen of *Cladophlebis* fragments. C, thin bedded siltstone with hopper casts, position within section is unknown, but is probably below the *Cladophlebis* bed.

silica and chondrite normalized La/Yb (Boynton, 1984) ratios than any of the Newark and Hartford basin flows (Fig. 5). Marzoli et al. (2004) argue that because the Tr-J boundary is so close to the base of the Orange Mountain Basalt, the lower unit of the Central High Atlas sequence should predate the boundary. They argue that this reflects a temporal sequence of magmatic events that can be used to correlate laterally. However, the North Mountain basalt of the Fundy basin is much closer geographically to the High Atlas basin than it is to the Newark or Hartford basins, and thus it might be expected to be more geochemically similar to Moroccan basalts. Indeed this is the case. The silica content and chondrite normalized La/Yb cn ratios of the North Mountain Basalt overlap those of the CHA lower unit (Fig. 35; Whiteside et al., 2005). Based on their own geochemical methodology, the North Mountain Basalt correlates with at least the lower basalt unit of the CHA series. However, the extinction horizon is represented in the strata immediately below the North Mountain Basalt, in the Partridge Island member that is in obvious conflict with their hypothesis.

Deenen et al. (2007), based on a reanalysis of date from the literature, argues that all of the basalts on the north shore of the Minas basin are more similar to those of the lower basalt formation in the CHA than they are to either the basalt of the south shore, the oldest basalts of the United States basins, or the intermediate basalt formation of the CHA. While the Lu/Hf and Y/Nb data from the north shore do tend to plot with the lower basalt formation, as indicated by Deenen (and Whiteside et al., 2007), there is no relationship between stratigraphic position of Y/Nb data and the three main basalt stratigraphic units on the north shore. In addition, two points plot in the south shore field but both come from the lower flow unit and are from the north shore. Sr data show a consistent trend with stratigraphy with the north and south shore being chemically homotaxial. The north and south shores are also physically homotaxial, as well. This suggests that the differences in Lu/Hf and Y/Nb data are a consequence of some process related to geography rather than temporal evolution of a single magmatic system. There is therefore no compelling reason to think the north shore flows are older than those on the south shore. For them to be so all of the flows on the north shore would have to be older than those on the south shore, but the two series would have to occupy mutually exclusive areas, because nowhere do they occur in superposition. To conclude, the simplest hypothesis consistent with the all of the data is that there is geographic variation in the Lu/Hf and Y/Nb data and at least within the North Mountain Basalt Lu/Hf and Y/Nb data do not track stratigraphy.

Whiteside et al. (2005, 2007, 2008) hypothesize that the high La/Yb cn ratios of the North Mountain and initial Central High Atlas basalt are a consequence of crustal contamination derived from long distance lateral transport from dikes emanating from then adjacent regions of the southeast United States and Senegal. This hypothesis is supported by magnetic anisotropy data of Ernst et al. (2003) and de Boer et al. (2003) who argue for long distance lateral transport of magma over 1000 km. We hypothesize that the Orange Mountain Basalt and its correlatives in the southeastern and northeastern United States are contemporaneous with the North Mountain Basalt of the Fundy basin and the lower and intermediate flows in the Argana and Central High Atlas basins, and that the lower La/Yb cn ratios are a consequence of erupting closer to the source of the magma, which de Boer et al. (2003) suggest was in the Blake Plateau region.

Marzoli et al (2004) and Knight et al. (2004) also argue that they can find no Jurassic strata below the Moroccan basalts. However, all the sections they figure and describe are deformed at the contact (c.f., Fig. 30). In all the cases we have examined, there is a unit virtually identical to the Partridge Island member below every basal basalt sequence (Olsen et al., 2002b; Whiteside et al., 2004; 2007) (Fig. 29, 30). This relationship suggests a simple test of their hypothesis. The prediction of the Marzoli et al. (2004) and Knight et al. (2004) hypothesis is that the High Atlas basalts were extruded *during* the carbon cycle perturbation reflected in the initial negative $\delta^{13}C$ anomaly of Hesselbo et al., 2002) and precludes the presence of a negative $\delta^{13}C$ anomaly beneath the basalts. These specific and distinct predictions are directly testable by additional sampling and analysis that we will discuss on the outcrop. While we find the correlation between large igneous provinces and extinctions compelling, we find there is still no direct evidence of a temporal sequence of events that allows or demonstrates causation.

It may be indeed that flood basalts and bolides produce similar effects on the Earth System. The Triassic-Jurassic boundary would seem to be an ideal venue for examining this possibility. Parsimony suggests, however, that one unifying rather than two different explanations should explain the same phenomena.

The basalt itself at Partridge Island shows some interesting features, In particular the lower flow has gabbroid layers in its upper part. Similar gabbroid layers produce zircons that are particularly useful in U-Pb dating because they crystallize in situ after the eruption and therefore directly date the flow, as opposed to magma chamber dynamics. At Digby, the lower flow has particularly thick and well-developed gabbroids which produced a $^{206}Pb/U^{238}$ age of 202±1 Ma by Dunning & Hodych (1992) and more recently an age of 201.27±0.03 Ma by Schoene et al. (2006). This age is indistinguishable from the marine Triassic-Jurassic boundary date of 201.58±0.17 (Schaltegger et al., 2008) and a new age of 201.7±0.6 Ma from British Columbia (Kunga Island) from just below the boundary (Pálfy et al., 2008).

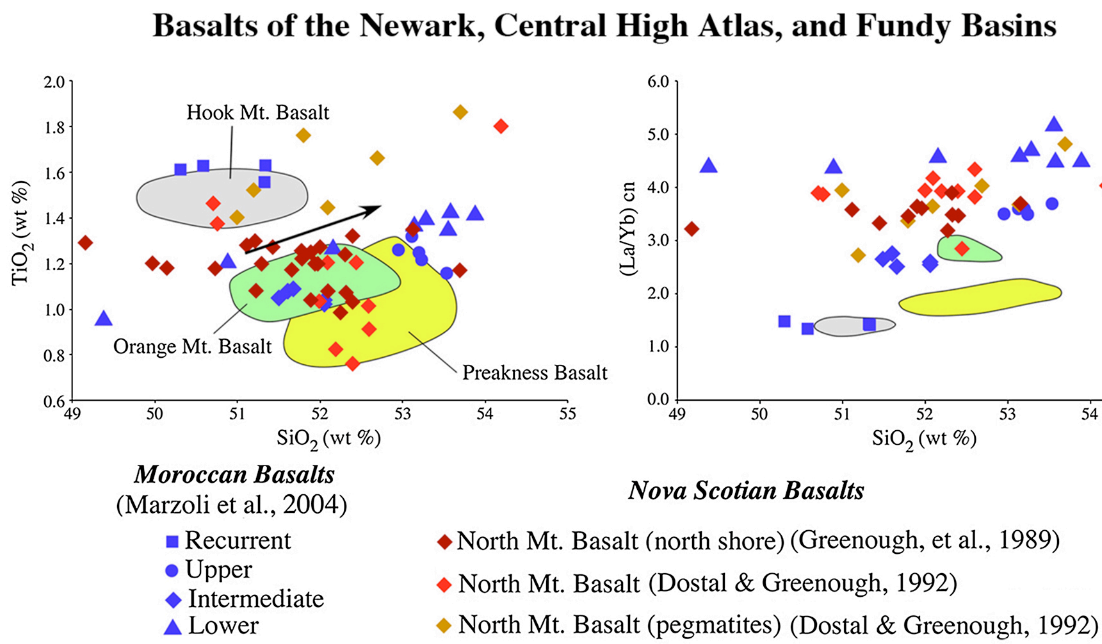


Figure 35: Aspects of the geochemistry of the Newark basin and Moroccan Basalts showing the similarity between the North Mountain Basalt (red, orange, and yellow) and the lower and intermediate basalts of Morocco (from Whiteside et al., 2005).

Return to vehicles. About 20 min to Stop 2.2

Retrace path back to Whitehall road	1.9 km
Head northeast on Whitehall Rd toward Lighthouse Rd	3.2 km
Continue on Main St	0.4 km
Turn right at 2 Island Rd	8.9 km
Arrive at parking for Wasson Bluff Protected area, Park	
Proceed down path to beach.	

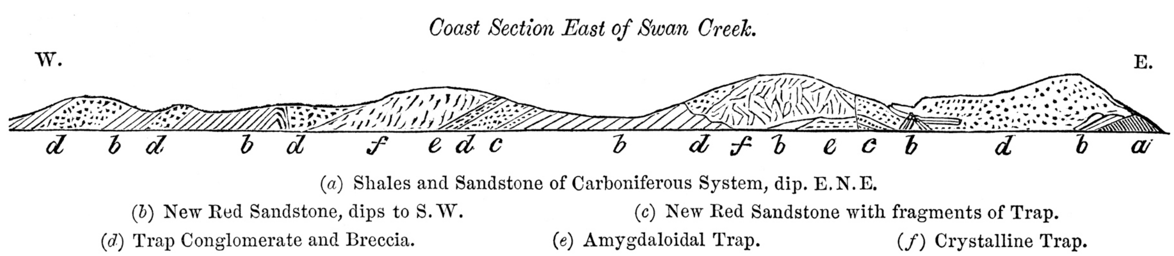


Figure 36: Dawson's (1855) depiction of the cliffs from Swan Creek (left) to Wasson Bluff (right). The Partridge Island member of the Blomidon Formation in contact with the North Mountain Basalt (Station 1 of Stop 5.3) is shown at the first "b" on the right. Station 12 (basalt cobble debris flows) is the first "d" on the left. Compare with the drawings in Nadon & Middleton 39.

Stop 2.2: Wasson Bluff Protected Area Traverse

45° 23.965'N 64° 19.392'W

Main Points:

Outcrop of the level of the end-Triassic extinction
Synfaulting sedimentation
Scots Bay Member lacustrine sequences
Earliest post-extinction tetrapod skeletal assemblages known; dinosaurs.

All collecting is prohibited by law at the protected Special Place. Do not use rock hammers anywhere on these outcrops, even if you see no fossils. Outside the protected site, it is still against the law to remove or disturb any fossil in the bedrock. All fossils in Nova Scotia have this legal protection.

Geologically, Wasson Bluff was first illustrated by William Dawson in 1855 in his depiction of cliffs east of Swan Creek to Wasson Bluff (Fig. 36). It was briefly described by Powers (1916) and Klein (1960) and developed a well-deserved reputation as a mineral locality, especially for zeolites (Dawson, 1891). 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), from which much of this is excerpted.

Wasson Bluff is the best place to see evidence for penecontemporaneous faulting and sedimentation during left-oblique slip of the Minas fault zone (Fig. 37). 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, 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, and hydroplastic slickensides probably formed in incompletely lithified sediments. Neptunian dikes are common in the North Mountain Basalt and reveal important kinematic information. 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. Over the past 30 years, both professional and amateur paleontologists have made significant discoveries of dinosaur and other fossils in these rich deposits (Fig, 38), the discovery and collection of which fossils was described by Olsen et al. (2005b). We shall examine all of these features on our trek along Wasson Bluff. The following discussion is based on Olsen et al. (1989, 2005b) and is tied to the sketches of the Wasson Bluff outcrops with a series of stations numbered from east to west (Fig. 39).

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. The Blomidon Formation is dipping towards the beach (i.e., southward)

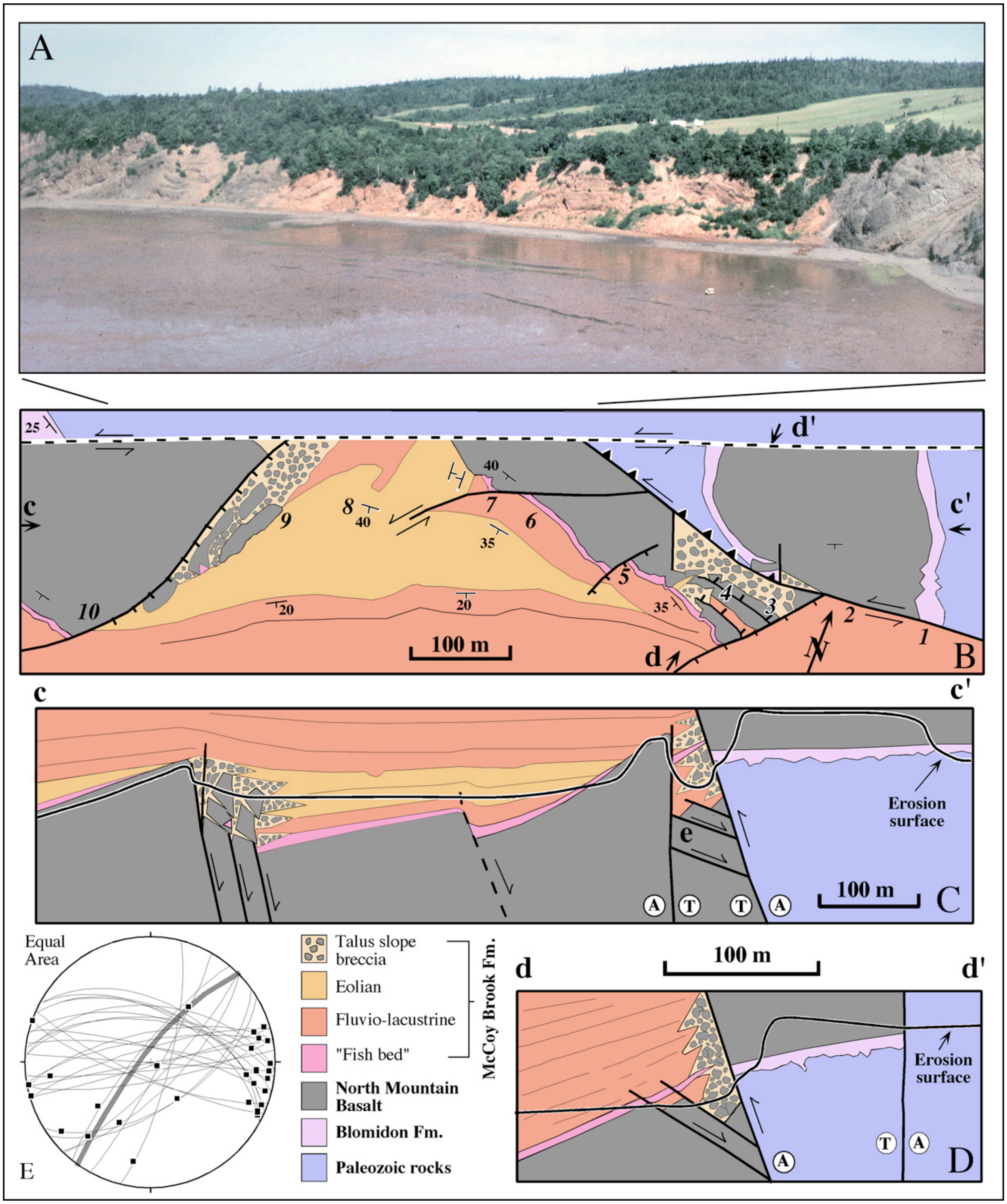


Figure 37: Wasson Bluff area, Stop 2.2: A, helicopter photo of middle segment of the Wasson Bluff transect; B, bedrock geological map of eastern 3/4 of Stop 2.2 transect (1-10 are Stations); C. cross-section c-c'; D, cross-section d-d'; E, equal area lower hemisphere stereographic projection showing attitude of faults (thin great circles), slickenlines (boxes), and average orientation of Neptunian dikes (shaded great circle) on west side of Wasson Bluff area, all consistent with northwest-southeast extension.

and consists mostly of grey, red and purple conglomerate and sandstone. The upper two meters, however consist of well bedded grey, red and purple sandstone and mudstone with some halite hopper cast-bearing grey claystones of the Partridge Island member. The contact between the Partridge Island member and the North Mountain Basalt is step-faulted. Hexagonal cooling joints are well developed immediately above the contact.

Station 2: The bulk of the North Mountain Basalt from here westward to the eastern sub-basin is mostly tectonized or "rubblized" basalt similar to what will seen at Stop 2.3. Schlische in Olsen et al. (1989) hypothesized that this extensive zone of "rubblized" basalt marks a wide, predominantly left-lateral fault zone. This same fault zone apparently places the McCoy Brook Formation against Carboniferous rocks to the east of these outcrops. Splays of this fault zone trending at a higher angle to the cliff face are present. The degree of rubblization increases towards these splay faults, and many are marked by narrow, chlorite-fiber slickensides. Again, we hypothesize that the "rubblized" basalt formed as a result of faulting under near-surface but not exposed conditions; the mineralized faults probably formed at greater depth as a result of burial.

In between some zones of "rubblized" basalt are interpreted as rotated basalt columns, much like those seen at Five Islands (Stop 2.3, Station 4). Stevens (1987) interpreted these rotated columns as portions of a dike. However, as at Five Islands (Stop 5.1), chill margins are absent, rotation of the columns is variable, and the sense of rotation is appropriate for down on the south motion of the bounding faults. Metamorphism and intrusive apophyses are also absent.

Station 3: A fault-bounded block of paleo-fault talus slope deposit marks the eastern end of the eastern sub-basin (Fig. 38). This sub-basin is easily recognized by the distinctive deposits consisting of angular clast-supported basalt breccia in a matrix of sandstone or mudstone. Large blocks of basalt 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. The matrix often yields abundant reptile bones. These deposits at Wasson Bluff are invariably associated with faults and have all the characteristics of talus slope accumulations (Tanner & Hubert, 1991).

Much like the talus slopes forming today on this beach, these McCoy Brook talus slopes contained many large empty spaces in which small animals could live. In order of abundance, the taxa found so far (Fig.38) include the protosuchid crocodyliform *Protosuchus micmac* (Sues et al., 1996), the very mammal-like trithelodont synapsid *Pachygenelus* cf. *monus* (Shubin et al., 1991), the sphenodontian *Clevosaurus bairdi* (Sues et al., 1994), a small probable ornithischian dinosaur resembling *Lesothosaurus* from southern Africa, and fragments of a small probable theropod dinosaur. Most material is dissociated, but some articulated material is present. 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. At this site aquatic fossils are completely absent.

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

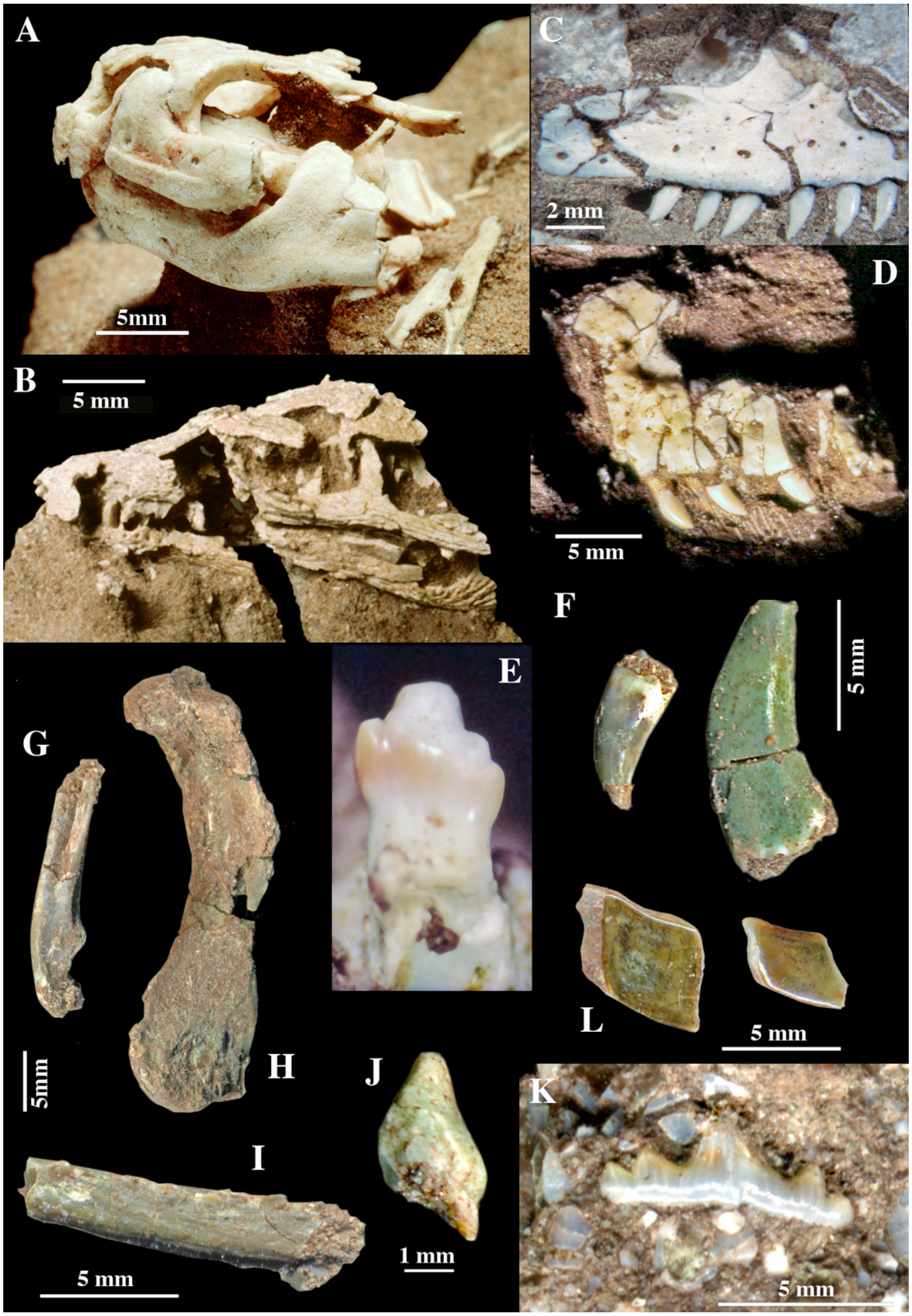


Figure 38. Small vertebrates from Wasson Bluff, Stop 2.2; A-B are from fluvial sandstone at Station 7, C-E, are from the talus slope breccias at Station 3, F-K are from the gravel bed of the Scots Bay Member in contact with the North Mountain basalt at Station 7: A, partial skull of sphenodontian *Clevosaurus bairdii* (type specimen); B, partial skull of protosuchian crocodyliform *Protosuchus micmac*; C, maxilla of undescribed sphenosuchian crocodylomorph; D, partial maxilla of *Protosuchus micmac*; E, molariform tooth of advanced cynodont *Pachygenelus* cf. *P. monus*; F, two teeth possibly of a small theropod dinosaur; G, left, ulna of an advanced synapsid, possibly *Pachygenelus*; H, humerus, possibly *Pachygenelus*; I, dorsal fin spine of hybodont shark; J, tooth of ornithischian dinosaur; J, tooth of hybodont shark, cf. *Hybodus* sp., K, scales of *Semionotus* sp.

the Blomidon Formation, in turn overlain by North Mountain Basalt (Fig. 37). 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 (Fig. 37). 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 lithologically-distinct Scots Bay Member. Lithologies typical of the Scots Bay Member are not present within the exposed dominoes to the north and the pink and white carbonate and green sandstone project on top of the units to the north. Therefore the sediments within these blocks should be basal-most 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 talus-slope deposits probably accumulated on the downthrown side of a normal fault. At the western end of this sub-basin, the talus slope deposits are faulted against North Mountain Basalt. This north-striking fault is probably a transfer fault because it terminates against the "reverse" fault, and dies out to the south in the zone of "dominoes".

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. 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. The dikes often follow pre-existing fractures, such as cooling columns in the basalt, or are sinuous. Schlische & Ackermann (1995) showed that the dikes in this region are preferentially oriented NE-SW in line with the regional extension direction.

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. The fault has offset the North Mountain Basalt—Scots Bay Member contact by approximately 25 m (Fig. 37). The fault zone is marked by a mineralized, slickensided plane with subhorizontal slickenlines. The basalt on either side of this plane is clearly rubblized, indicating the fault origin of the "rubblized" basalt. 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.

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 (Figs. 37, 40). Stratigraphically-higher strata onlap boulders of North Mountain Basalt, which in turn rest against a high-angle wall of massive vesicular basalt. The western edge of the basalt wall is truncated by onlapping fluvial sandstone. Thin layers of basalt rubble are interbedded in the sandstone adjacent to the basalt. It is possible that the surface of the basalt was terraced by the lake that deposited the deeper water parts of the Scots Bay Member, producing the evident paleorelief.

Onlap of the Scots Bay Member onto the basalt is marked by a gravely mudstone very rich in disarticulated fish and small tetrapod bones (Fig. 38). The most abundant faunal elements are disarticulated remains of *Semionotus* sp. that can comprise a coquina of scales. Second are

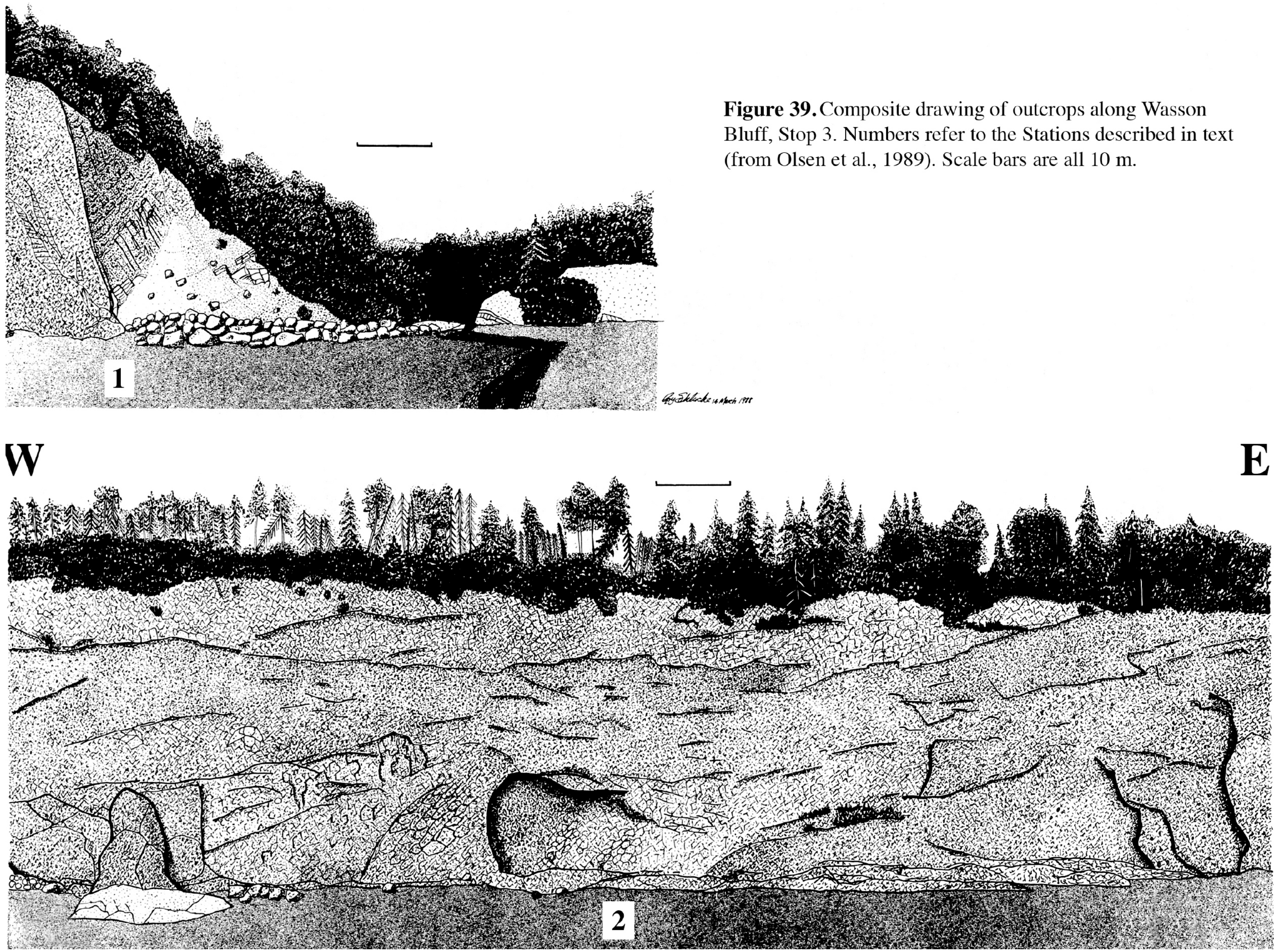
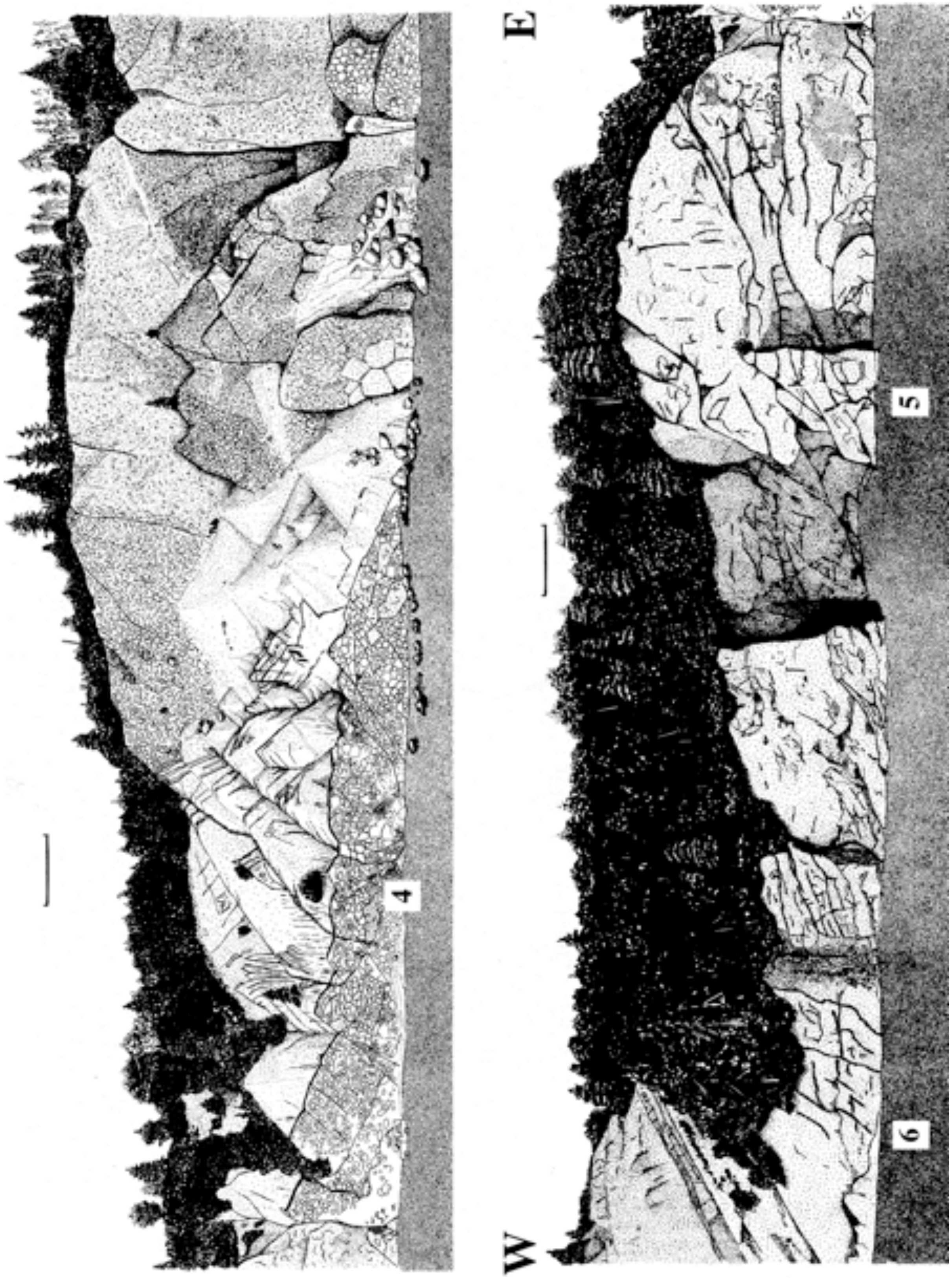


Figure 39. Composite drawing of outcrops along Wasson Bluff, Stop 3. Numbers refer to the Stations described in text (from Olsen et al., 1989). Scale bars are all 10 m.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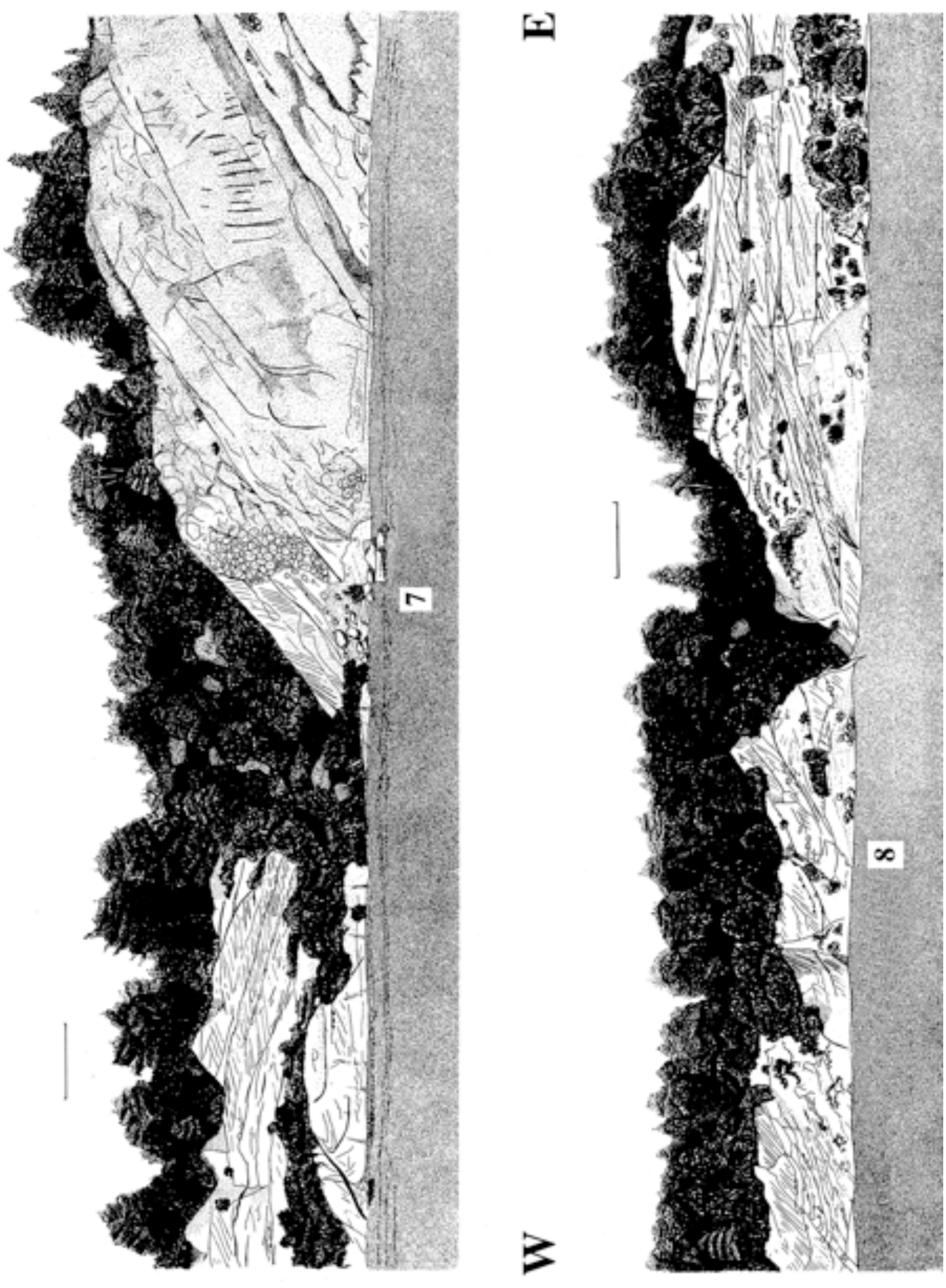

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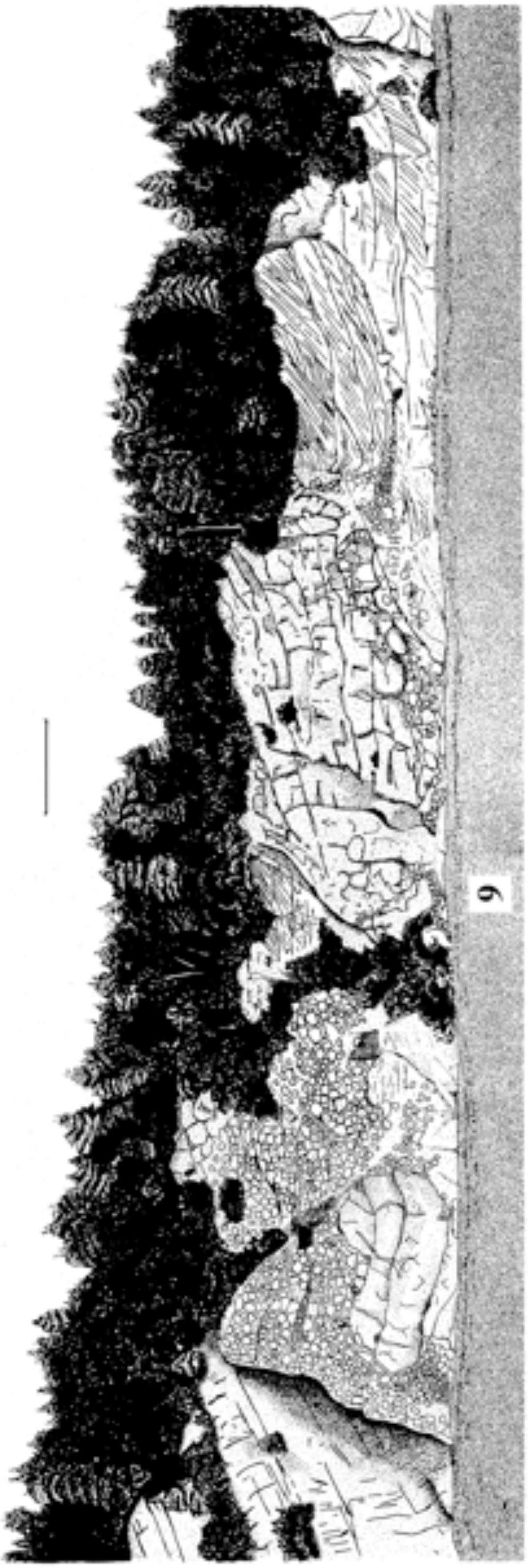

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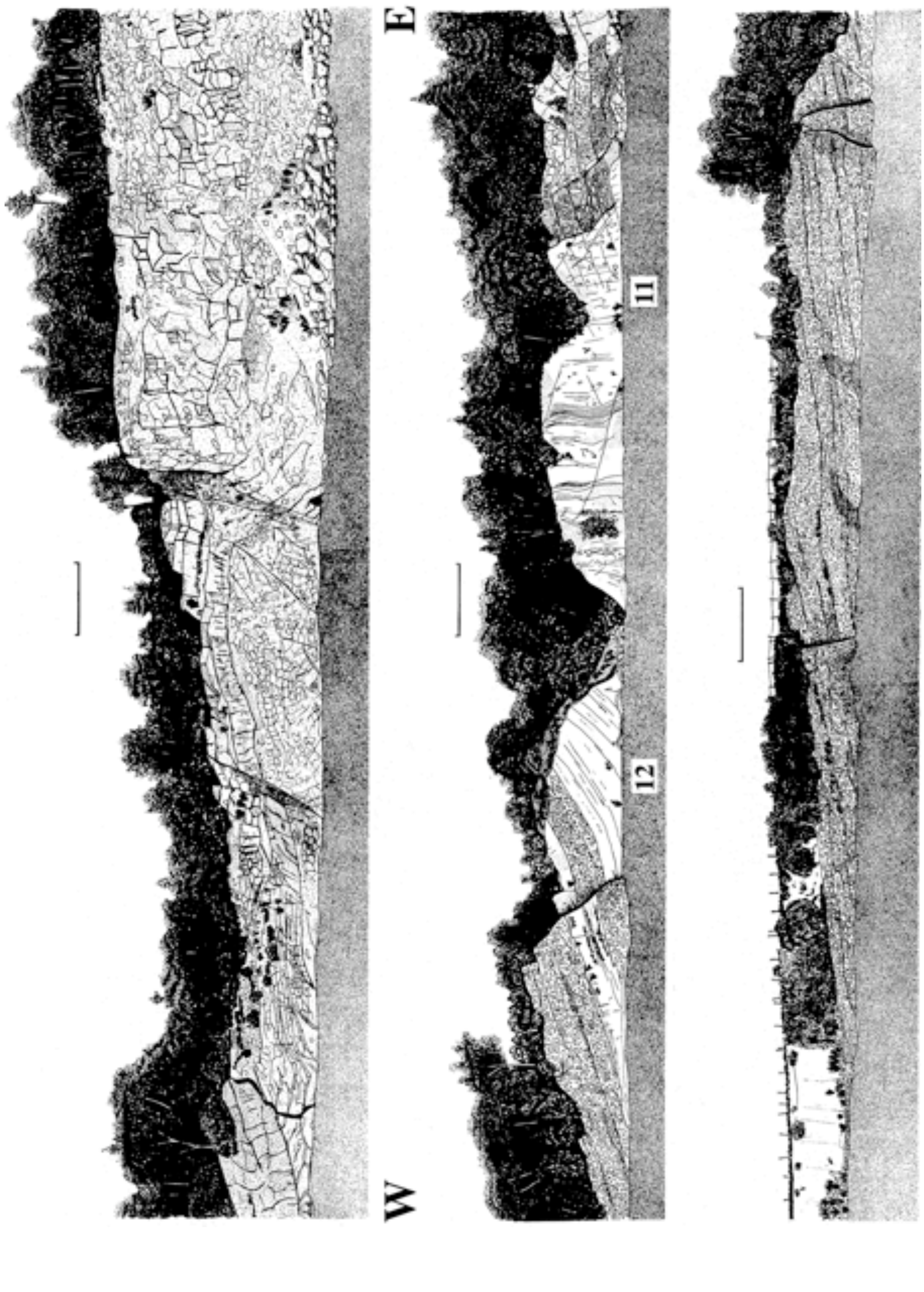
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teeth and spines of a hybodont shark close to *Hybodus*, followed by scale and skull bones of an undetermined redfieldiid fish. Tetrapod bones are surprisingly common; no particular taxon is seemingly dominant. Isolated elements include protosuchid girdle elements and osteoderms, small theropod dinosaur teeth similar to *Syntarsus* (or *Coelophysis*), small ornithischian teeth, vertebrae, and limb elements, and probable trithelodont postcranial elements. This gravelly unit is a wave-sorted lag that accumulated between cobbles along the Scots Bay lake shore. Again, as was the case for the carbonate sequence of the Scots Bay Member at Stop 4.1, these lakes were too shallow to preserve significant organic matter.

The upper portions of the lacustrine sequence pass upward into red mudstone and sandstone beds. The latter contains well-preserved reptile remains, some of which are articulated, but as yet no aquatic elements. 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 most common taxon is the lizard-like sphenodontian *Clevosaurus bairdi* (Fig. 38). The second most common taxon is *Protosuchus micmac*, represented mostly from abundant isolated elements but an articulated if fragmentary skull has also been found (Fig. 38). Third in abundance are isolated elements of the trithelodont *Pachygenelus* cf. *monus*. Last are possible dinosaurian remains, the most notable is a possible articulated yearling prosauropod skeleton.

Figure 40: Onlap of the Scots Bay Member (white and pale purple) of the McCoy Brook Formation (Station 7, Stop 2.2) onto a possible wave cut notch on the upper surface of the North Mountain Basalt as it looked in 1987. Note that the Scots Bay Member is covered by a paleo-talus cone that contains a lag of fish and other bones.

The McCoy Brook assemblages occur directly on top of the North Mountain Basalt, which based on geochemistry is the temporal equivalent of the Orange Mountain Basalt, and therefore the assemblage is plausibly correlated with the Midland Formation of the Culpeper basin, the Feltville Formation of the Newark basin and the Shuttle Meadow Formation of the Hartford basin. Based on the geochemical similarity of the North Mountain and Orange Mountain-Talcott basalts (Puffer & Philpotts, 1988) and the cyclostratigraphy of the upper Blomidon and lower Scots Bay Formation, the McCoy Brook assemblages is less than 100,000 years younger than the end-Triassic extinction event (Olsen et al., 2002b, 2003b; Whiteside et al., 2005). The extinctions must have taken place prior to this time suggesting an abrupt extinction event.

Station 8: This area lies near the center of the triangular western sub-basin and the cliffs in front of us (Fig. 37) are representative of most of its fill. 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 which yield paleo-wind directions towards 241°. 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 (Fig. 37), 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). According to these authors, 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 barchan dunes and barchoid 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.

Less impressive eolian sandstones have been reported from age equivalent strata of the Cass Brook Formation of the Pomperaug basin between the Newark and Hartford basins, equivalent to the Shuttle Meadow Formation, and from the upper Shuttle Meadow Formation itself (LeTourneau & Huber, 2006). It appears that although the strata deposited within a few hundred thousand years of the end-Triassic extinction event were characterized by a general increase in humidity (as witnessed by the Scots Bay Member, pronounced arid episodes still occurred, suggesting perhaps an amplification of the expression of both extremes of the hydrological cycle, perhaps due to elevated CO_2.

Directly in front of us is the site of the discovery of a bone bed with numerous articulated to disarticulated sauropodomorph “prosauropod” dinosaurs, comprising the largest assemblage of prosauropods in North America. Intriguingly, although prosauropods are abundant from strata fro mid to high latitudes in both hemispheres during the Late Triassic, there are no Triassic prosauropods at all from North America, even from the richly fossiliferous Chinle Group of the southwestern United States. However, by 100 – 300 ky after the end-Triassic extinction event, they were abundant in North America as witnessed by this locality and footprints (*Otozoum*) in more southern CAM basins and the Wingate Formation of Glen Canyon Group.

Finally, looking south, laterally persistent beds of apparently ripple cross-laminated, water-laid, reworked eolian sandstone are found in the tidal flat at the top of the eolian sandstone wedge in the main half graben. These sandstones contain poorly preserved but abundant brontozooids including *Eubrontes* (*Anchisauripus*) cf. *sillimani* and *Batrachopus* sp. tracks.

The vertebrate assemblages from Wasson Bluff are critical to the debate on the structure and origin of the Triassic-Jurassic mass extinctions. Compilations of latest Triassic assemblages from other portions of the Newark Supergroup and other parts of the world (Olsen and Sues, 1986; Olsen et al., 1987; Benton, 1995) demonstrate that all of the families present in the Early Jurassic McCoy Brook assemblages were already present earlier in the Triassic. However, despite the fact that the Nova Scotian post-end Triassic extinction event assemblages come from a variety of facies, from fully aquatic to fully terrestrial, and the fossils are very common, the dominant late Triassic families of reptiles and amphibians are absent. This is also true of intra- and post-extrusive deposits from the rest of the Newark Supergroup. Thus, the McCoy Brook assemblages represent survivors of the massive Triassic-Jurassic extinction event, in which about 50% of all terrestrial vertebrate families went extinct.

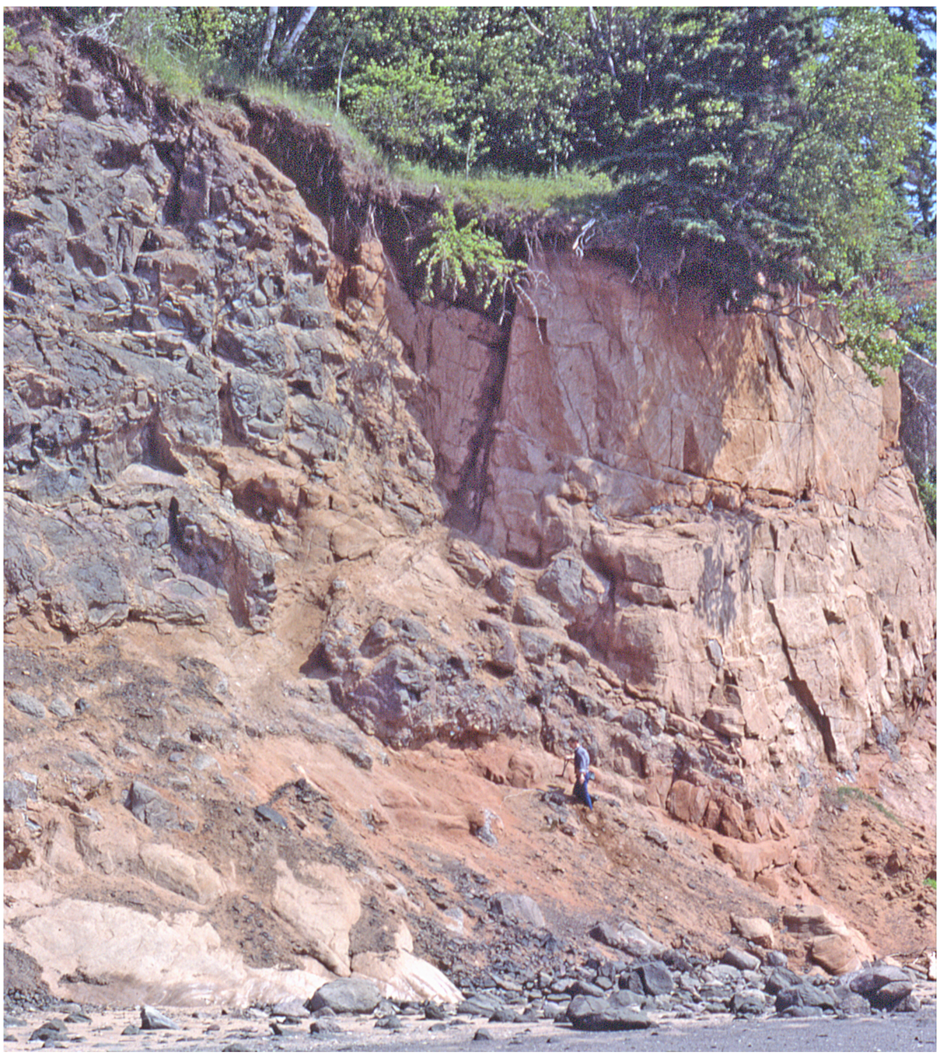


Figure 41: Paleo-talus cone in cross-section at station 9 (Stop 2.2) as it appeared in 1986. On the left is a fault block of North Mountain basalt with a buttress unconformity with younger eolian McCoy Brook Formation on the right, containing a well defined talus cone of basalt cobbles derived from the immediately adjacent paleocliff. Some of the blocks have rolled tens of meters away over the dune surface, while others have moved only slightly away from the cliff.

Station 9: A complex mix of very large basalt blocks (+5 m), sandstone and talus breccia marks the west end of this sub-basin, just east of its bounding fault. A beautiful outcrop section of a paleotalus cone existed here in previous decades (Fig. 41), but most has eroded away. A fragmentary but partially articulated prosauropod dinosaur skeleton includes a partial gastralia basket and associated group of gastroliths. This is the only North American prosauropod known with gastroliths. They are especially interesting because clasts of this type are not found in the Wasson Bluff outcrops, so the dinosaur must have traveled some distance to get them. Elements from this specimen were included in a study of dinosaurian gastroliths (Whittle & Onorator, 2000), and a sphenodontian mandibular ramus located several centimetres from the gastrolith pile has been used to suggest prosauropods may have been omnivorous (Barrett, 2000). It should be noted, however, that phosphorus is limiting in most environments and it is normal for otherwise herbivorous tetrapods to chew or ingest bones and bone fragments. The prosauropod remains were badly broken prior to fossilization and deformed from syndepositional faulting that commonly affects vertebrate fossils from this formation. A structural inference can also be based on the gastroliths and the skeletal deformation; faulting took place under low confining pressures in the relatively incompetent, probably incompletely lithified sandstones. None of these intra-dinosaur faults are mineralized. Parts of this specimen are on display at the Fundy Geological Museum.

The southeast corner of the large multi-flow basalt block between sandstone on the east and massive basalt flows on the west is notched and filled with Scots Bay Member, containing its characteristic fauna, apparently in gross unconformable contact. We interpret this notch as a cavity produced by (wave?) plucking of a basalt boulder followed by onlap of the Scots Bay. The Scots Bay Member is truncated to the immediate southeast by a healed fault, the south side of which is orange sandstone and basalt breccia. Again this series of contacts shows that faulting, erosion of the uplifted blocks, and sedimentation were contemporaneous.

The normal fault (as revealed by slickensides) at the northwestern edge of this sub-basin is typical of faults at Wasson Bluff. It consists of an upward-widening fault zone of (basalt) breccia in a matrix of sandstone and mudstone. The degree of brecciation and volume of sedimentary matrix increase toward the fault. This fault breccia is welded to the hanging wall and partially to the footwall, and is cut by a narrower zone of red fine-grained material, which is probably gouge, but resembles the infill of Neptunian dikes. This red material is then cut by narrow, mineralized, slickensided surfaces. We hypothesize that the matrix-rich breccia formed in a wide fault zone that was open at the surface. The red gouge and mineralized slickensides formed at greater depths as a result of additional burial. It therefore appears likely that fault zones narrow with depth (within the brittle field), with cataclasis becoming more confined, and with eventual mineralization.

Station 10: Westward of the sub-basin are extensive exposures of middle and upper North Mountain Basalt. Several thin flow units are overlain by the thick and massive upper flow sequence of the basalt, all of which dip to the southwest. Some evidence of faulting during extrusion of the North Mountain Basalt includes a left-lateral offset in a wrinkle or flexure at the top of one thin flow, filled by the succeeding next thin flow. The cooling rims in the infilling flow suggest that the offset occurred before the extrusion of the upper flow. Thin, dark, vesicular zones are segregation veins that are enriched in silica and sometimes zirconium. Zircons from segregation veins of the North Mountain Basalt near Digby, have yielded an ^{206}PB/^{238}U age of 201.27±.03 Ma (Schoene et al., 2006) which is nominally slightly younger than a 201.58 ± 0.17

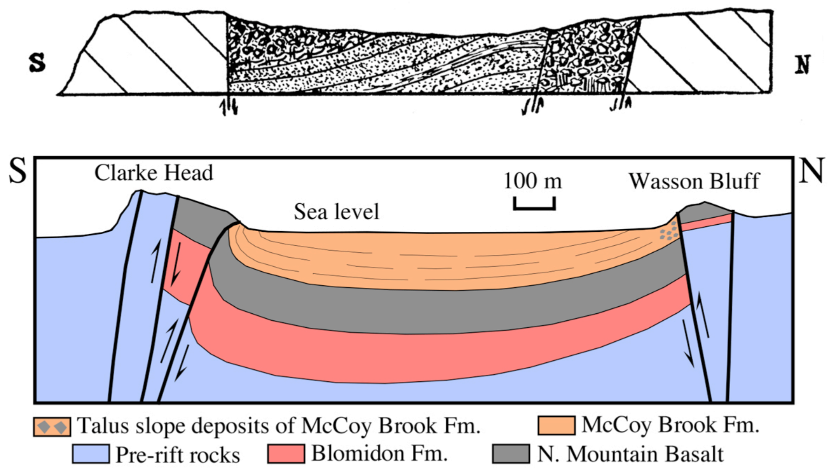


Figure 42. Cross-section from Wasson Bluff to Clarke Head. Above, Powers (1916) interpretation in which basaltic agglomerate lies above Wolfville Fm.. Below, interpretation of Withjack et al. (1995) in which the stratigraphy is strongly modified by syn-sedimentary tectonics and there is a thick post-basalt unit (McCoy Brook Fm.)

Ma age from zircons from a tuff from marine strata very close to the end-Triassic extinction event, demonstrating how close in time the oldest CAMP basalts and the extinction even were.

Hexagonal cooling joints are well developed in some of the flow units at the western side of this station. Many of these joints have served to localize the Neptunian dikes. In one example described and figured by Schlische & Ackerman (1995), we see that the cooling joints oriented normal to the regional extension direction formed the widest Neptunian dikes.

In this same area, a sinuous 20- to 30-cm-wide vesicular basalt dike with clearly chilled margins cuts a flow. Its relationship to the surrounding units is completely obscure and it may represent no more than an injection of the last fluid material from the interior of a cooling flow into its already chilled upper surface.

Station 11: 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 (Figs. 37, 39). The fault zone between the vertical McCoy Brook Formation and the basalt contains phacoids of purple basalt gravel. A less-disturbed, similar-appearing gravel is present in the adjacent mud flat directly along strike, in conformable contact with the North Mountain Basalt, and is apparently the local expression of the Scots Bay Member. 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. Its stratigraphic position in the section at Station 7 seems to be just above the fluvial sandstone capping the Scots Bay Member. Thus far it has produced only very rare ostracodes, conchostracans, and *Semionotus* scales. From this station, and perhaps as far west as Swan Creek, is another small sub-basin developed on the syndepositionally faulted and tilted North Mountain Basalt with its own distinctive local facies of the McCoy Brook Formation.

The small creek that comes down from the north at this station follows the strike of the steeply dipping McCoy brook until it crosses a fault and passes into the Blomidon Formation. The Blomidon Formation is much thicker here than at Stations 1 and 2 and looks much more like it does at Partridge Island, demonstrating thickening of the formation towards the Clarke Head area (see below).

Station 12: 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. The clasts of the matrix-supported flows consist exclusively of North Mountain Basalt. Individual debris flows are only decimeters thick (J.P. Smoot, pers. com.). Tanner & Hubert (1991) have further described this distinctive but unfossiliferous debris flow sequence. 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. Previously thought to be lacking in vertebrate remains a partial, fragmentary but articulated *Protosuchus* was found in 2007 on a small cobble on the beach adjacent to the outcrops (by Gustaf Olsen). 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 shows a reverse fault throwing North Mountain Basalt over McCoy Brook Formation. Superficially, it appears as though the North Mountain Basalt is in conformable contact with underlying red beds, contributing, no doubt to the misinterpretation that the red beds at Wasson Bluff were Blomidon Formation and the basalt at Wasson Bluff (and McKay Head) were interbedded between the Blomidon and Wolfville Formations (e.g., Klein, 1962). Another fault to the south separates North Mountain Basalt from a tectonic mélange of Palaeozoic rocks, which include boulders of granulite-grade mylonite (Gibbons et al., 1996), developed within the most outboard exposed fault segment of the Minas fault zone. While Gibbons et al. (1996) ascribe the deformation to successive faulting events, the gypsiferous megabreccia exposed at Clarke Head finds a remarkably close match in breccias seen in outcrop along the shores of the Isles de Madeline in the Gulf of St. Lawrence. In this area, the breccias are due to salt diapirism, another deformational process that should be added to the mix of processes acting along this remarkable fault zone.

An interpretive cross section from here to Clarke Head is shown in Nadon & Middleton 42 (from Withjack et al., 1995). The synclinal shape of the overall structure is supported by measurements of McCoy Brook bedding attitudes in the beach. 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.

Paleozoic sedimentary rocks to the east of the faults shown at Clarke Head in Figure 42, are highly deformed and a gypsum cemented melange is present. This is probably the remains of a Mississippian Windsor Group salt and gypsum diapir that invaded the fault zone and intensely deformed the surround units. How much of the deformation seen along the Minas fault zone could be due to Triassic-Jurassic salt tectonics is under-appreciated and virtually unexplored. Windsor salt was presumably remobilized during Late Permian and Early Mesozoic extension, and it certainly might explain some of the anomalous Triassic and Jurassic structures along this coast.

Time permitting, we will be leaving Stop 3 by a road paralleling Swan Creek. As we follow Swan Creek inland, we will pass outcrops of the Scots Bay Member with fragmentary *Semionotus* and coprolites on the west bank of the creek where it intersects the North Mountain Basalt. This way of leaving is approximately 2.8 km south west of where we parked.

Return to vehicles (about 50 min to Stop 2.3).

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
Park at beach area parking, proceed north down to beach	

Stop 2.3: Five Island Provincial Park

45° 23.671'N 64° 3.710'W to 45° 23.005'N 64° 02.211'W

Main Points:

- Spectacular overview of Triassic-Jurassic transition
- Structurally condensed section near Minas Fault zone
- CAMP basalts
- Jurassic McCoy Brook fluvio-deltaic elements

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 (Figs. 18, 27). The Minas fault zone has had a complex tectonic history, and the sections are much more structurally complex, the section being unique in having 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.

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 (Fig. 43). 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.

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 (Nadon & Middleton 44) and several meters of interbedded fluvial sandstone and pebbly sandstone and conglomerate (Hubert & Mertz, 1980, 1984). Although these strata were originally thought to be part of the Wolfville

Formation, mostly because they were coarse grained (Klein, 1962), they are interbedded with facies typical of the lacustrine portions of the Blomidon Formation, relationships that can best be seen further to the east at Lower Economy (Fig. 44). At the latter locality, the Red Head beds overly the Wolfville Formation (TS II) with a profound angular unconformity, a relationship we will discuss at Station 6.5 According to Hubert & Mertz (1984), the eolian strata were deposited as barchan-type dunes by winds bowing to the southwest (254º) winds. The direction of these winds is consistent with the Northeast Trade Winds and a location near the Tropic of Cancer (N 23.5°) in Triassic time.

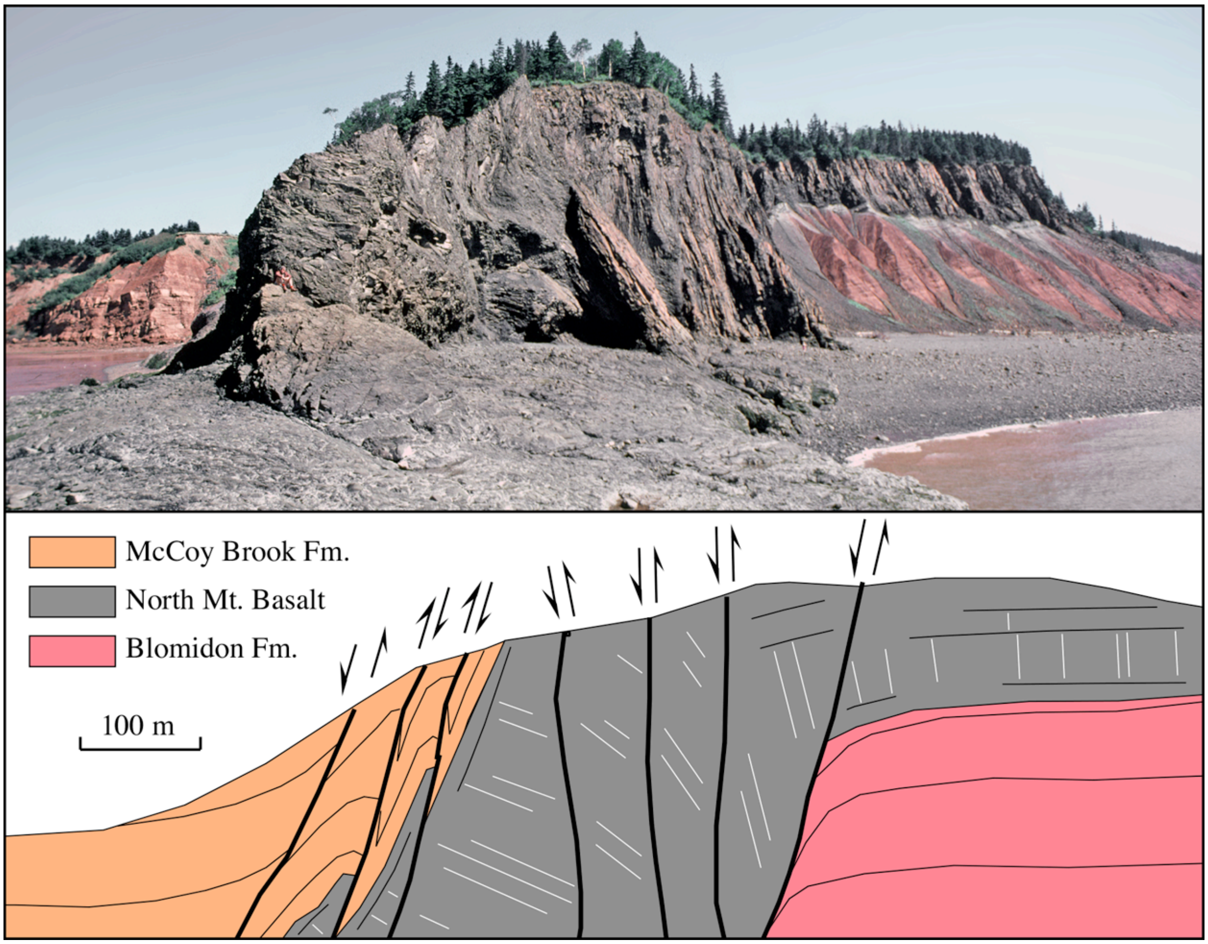


Figure 43: View from Station 4, Stop 2.3: above, wide-angle view to the NE from the Old Wife showing Blomidon Formation conformably overlain by North Mountain Basalt on the right, down faulted North Mountain Basalt straight ahead; and down faulted and reverse faulted McCoy Brook Formation on left. Below, cross-section (from Withjack et al., 1995).

Very similar sequences occur at the base of TS III over the western Moroccan CAM basins (Et-Touhami & Olsen, 2003). These strata comprise the Tadrart Ouadou sandstone in the Argana basin and the uppermost part of the Oukaimeden Sandstone in the High Atlas area, and are thought to be the main reservoir for the producing Meskala gas and condensate producing field in the Essaouira basin. However, the polarity stratigraphy of the Blomidon Formation of the Fundy basin and the Bigoudine Formation of the Argana basin shows unequivocally that the eolian sandstones are not the same age in the various basins (Fig.45). Instead, these eolian

Figure 44: Red Head member of the Blomidon Formation: left, eolian strata just east of fault separating the Whitewater member from the Red Head member at Station 1 of Stop 5.1, at Red Head; right, lacustrine strata typical of Whitewater member interbedded with eolian and fluvial strata of Red Head Member at Lower Economy.

sandstones occur in a homotaxial position probably controlled by the proximity of an unconsolidated sand source and a high water table to preserve the toes of the dunes, all within the context of the arid subtropics. The Red Head Member is equivalent in time to the wetter-looking members Q, R, and S of the Passaic Formation Newark basin at about 214 Ma (based on the NBAGPTS) that does however fall in the dry phase of a 3.5 million year cycle. In contrast, the Tadrart Ouadou Member correlates to about Member C of the Newark basin at about 217 Ma, in a dry phase of a 1.75 and a 3.5 million year cycles. It thus appears that climate plays an important role in localizing the dune sequences, within the appropriate tectonostratigraphic context.

Walk west-northwest about 150 m around the sandstone point.

Figure 45: Interbedded eolian and fluvial units of Tadrart Ouadou Member of the Bigoudine Fm. of the Argana basin north of Argana, Morocco. Height of forests in upper right corner is about 1.5 m. View is to the northwest, and paleowinds blew here to the southwest.

Station 2. Lacustrine middle White Water Member and dissolution features: A southwest striking, northwest dipping fault separates the Red Head member from the middle Whitewater member of the Blomidon Formation (Fig. 46). 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 to the southwest at Cape Blomidon and the GAV-77-3 core. There are two fish-, conchostracan-, and ostracode-bearing Van Houten cycles present, the upper cycle has a highstand interval that is distinctly laminated with purple, yellow, and green bands. This upper laminite is brecciated downward to contorted sands and siltstones, and overlain by a granule-bearing conformable sandstone bed. Strata below the lower fossiliferous laminite are deformed by numerous bed-specific normal faults of various sizes. This is the same deformation zone and distinctive lithologies we saw at Stop 4.2, interpreted by Olsen et al. (1989) and Ackermann et al. (1995) as a result of salt (halite) dissolution. Tanner (2006) however interpreted the deformation as due to a strong synsedimentary seismic event, namely the impact of the bolide that produced the Manicouagan structure. We believe 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. In addition, based on magnetic polarity stratigraphy of the GAV-77-3 core, these strata correlate with the E17r reverse polarity magnetic chron of the Newark basin time scale NBAGPTS (Kent & Olsen, 2000) while Manicouagan is of normal polarity. 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 46: Penecontemporaneous deformation attributed to evaporite (halite) dissolution at Station 2, Stop 5.2. *SC*, marks the level of salt collapse beds that is the same as at Stop 4.2 at 2570-2636 m. Domino-style faulting at lower right is also attributed to salt dissolution.

Station 3: 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. Some of the cycles are especially well developed. The majority of the section is of Rhaetian age. Hundreds of mostly small-displacement normal faults subtly cut the section, 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. The upper beds of this interval have produced blackened grains of *Corollina* pollen, and elsewhere the uppermost Blomidon typically has grey and black palynomorph-bearing mudstones that often weather white. This is the Partridge Island member marking the base of TS-IV. However, at this locality the thickness of the Partridge Island member is clearly enhanced by metamorphism from the overlying North Mountain Basalt, a pattern seen in several other areas including Grand Manan Island. Based on the occurrence of the palynological transition at Partridge Island and the lateral continuity of the member, the end-Triassic extinction almost lies within the upper meter of the white layer.

Neither the stratigraphy of the Partridge Island member nor the visible manifestation of the metamorphism measurably change laterally here. 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. A less metamorphosed section of the Partridge Island member can be seen nearby on Pinnacle Island to the southwest, where typical black and gray mudstones are interbedded with red mudstones and gray sandstone within a meter of the North Mountain Basalt contact.

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.

Station 4, 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 (Fig. 43). Further to the northwest and along strike of the low-tide peninsula that supports the sea stack known as the Old Wife, a succession of high angle faults and basalt rubble zones are present between the faults and rubble zones are bands of horizontal hexagonal columns. Like all the surfaces of the individual clasts in the basalt rubble zones, which are monomict, all surfaces of the columns are slikensided. This zone of horizontal columns was originally interpreted as a dike that fed the North Mountain Basalt (Powers, 1916; Stevens, 1987). Olsen et al. (1989) and Withjack et al. (1995), however, argue that this zone is simply a rotated zone of the middle part of the North Mountain Basalt, the thinner flows of which commonly have columnar jointing. The rotation follows the pattern seen in the less deformed, lower flow of the North Mountain Basalt to the east. There is no evidence of metamorphism in the surrounding basalt, and there is a complete lack of intrusive apophyses into the surrounding basalt or sediment, as seen in undoubted dikes intruding or feeding basalt flows (e.g., Philpotts & Martello, 1986; Olsen et al, 2003b). All the contacts are instead brittle faults with slikensides.

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. Proceed northwest onto the beach and walk along the beach contact with the North Mountain Basalt.

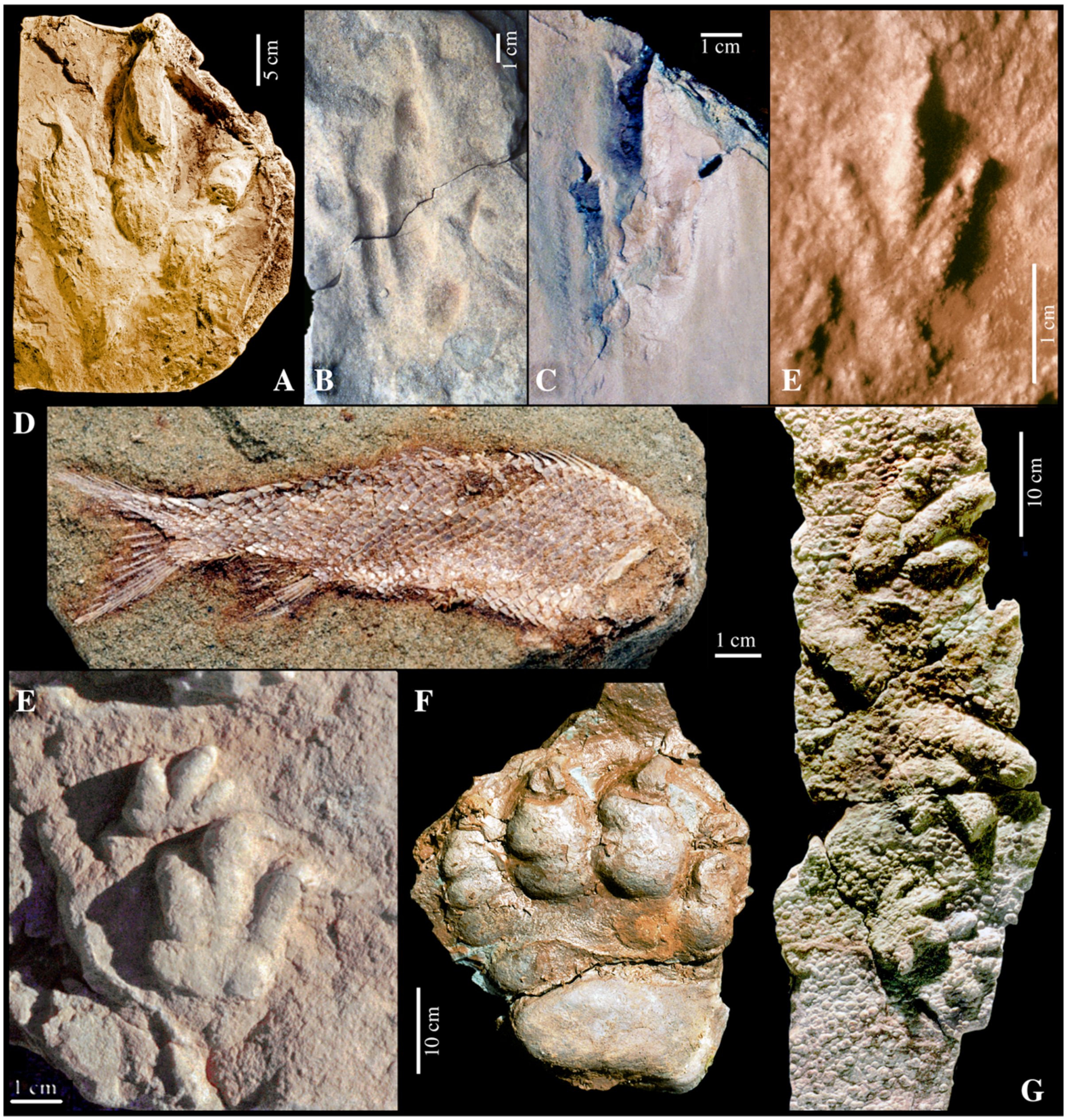


Figure 47: Representative vertebrate fossils from the McCoy Brook Formation of Five Islands (A-D), and near-by Blue Sac and McKay Head (E-G): **A**, *Eubrontes* cf. *giganteus*; **B**, *Eubrontes* (*Anchisauripus*) cf. *hitchcocki*; **C**, *Eubrontes* (*Grallator*) cf. *cursorius*; **D**, *Semionotus* sp. (photo by Heinz Wiele); **E**, cf. *Eubrontes* (*Grallator)* sp.; **F**, *Batrachopus deweyii,* Eldon George collection; **G**, *Otozoum* cf. *moodii*; **H**, *Anomoepus scambus*.

Station 5, Contact with the McCoy Brook Formation: All along this contact, depending on the tide and sand movements, we will be able to see steeply dipping and sheared red beds of the McCoy Brook Formation in contact with mostly polymict breccia of the North Mountain Basalt, often with red matrix. Outcrops on the tidal flat have rapidly shallowing dips away from the basalt.

The contact between the McCoy Brook Formation and North Mountain Basalt on the shoreline is partly sheared (Fig. 43). At the contact he 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. 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 gradational passing from vesicular basalt upward into a polymict basalt breccia, a polymict cobble conglomerate and finally a red siltstone with basalt clasts. The contact therefore seems a normal sedimentary one, albeit somewhat sheared.

Examination of the fault surfaces reveals at least two sets of slikenlines. One set is oriented essentially parallel to dip, another is clearly strike slip, in some cases nearly horizontal. In some cases the strike slip slikenlines seem superimposed on the dip-slip slikenlines but this relationship is far from unambiguous.

The northwestern block of basalt has some very interesting internal contacts. There are at least two internal flow contacts with chilled margins that have nearly horizontal dips. The upper one has lovely pipe vesicles stained with malachite. However, these flows pass seamlessly both laterally and vertically into breccia and the bedding attitude of the flows seems irreconcilable with the contact with the McCoy Brook Formation. 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.

There are no indications of the Scots Bay Member at the contact here, however, nearby, P.E.O. has observed typical purple limestones and green sandstones in contact with the upper surface of the North Mountain Basalt on Long Island to the southwest.

In summary of tectonic inferences from Stations 4-5, the large-scale sense of motion from looking east to west from Station 4 is clearly down to the northwest consistent with the sense of rotation of bedding and fractures, as well NW-SE early Mesozoic extension. However, at the smaller scale there are reverse faults that indicate later shortening along the same fault zones. This 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). In addition, there is evidence of some syn-basalt tectonism, which consists of the discrepant bedding attitudes of the basalt flows seen at this station. One way to account for these bedding attitudes is that the flows were rotated prior to the deposition of the overlying McCoy Brook Formation, implying syn-extrusive faulting. As there is very strong evidence for syndepositional faulting effecting McCoy Brook sedimentation at Wasson Bluff (Stop 5.3), syn-basalt faulting at this location would not be surprising. This is also consistent with the apparent lack of the Scots Bay Member here and its widespread presence elsewhere including on the adjacent islands. It may not have been locally deposited because this portion of the North Mountain Basalt may have remained above the level of the lakes that deposited the Scots Bay Member.

Proceed northwest, along the long cliff outcrops of the shallow-dipping McCoy Brook Formation.

Station 6, Cliffs and beach outcrops of the McCoy Brook Formation: We will stop at 5 spots along the outcrop, derived from Olsen et al. (1989).

Station 6.1: 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. Centimeter- to decimeter-scale beds high in the sequence have dinosaur footprints, and were the source of the very first found in fallen blocks in the 1970's.

Station 6.2: Location of common rock falls from fish-bearing thin-bedded sandstone and mudstone in cliff. Fish-bearing unit consists of a climbing-ripple cross-laminated or convolute-bedded sandstone overlying a laminated, purple-red claystone. Complete *Semionotus* sp. (Fig. 47) occur densely packed at the top of the clay and in the base of the sandstone. The underlying claystone also has some partial to complete fish and coprolites. The fish-bearing sandstone represents a single sedimentation event that entombed the dead fish after a mass kill. The articulated preservation resulted not from burial in beneath anoxic waters of a deep meromictic lake, but rather from catastrophic burial by silts and sand in a very shallow lake, after which the oxidizing, ecologically efficient shallow ground water microbial community respired all the organic material present, without disrupting the bones.

Station 6.3: 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 and provide biostratigraphic evidence of post-Blomidon age of these strata (Olsen & Schlische, 1990). Ichnotaxa identified, using the nomenclature of Rainforth (2005) so far include the brontozoids (grallatorids of older literature=theropod dinosaur tracks) *Eubrontes giganteus*, *Eubrontes* (*Anchisauripus*) *E. parallelus*, *Eubrontes* (*Anchisauripus*) *E. sillimani, Eubrontes* (*Anchisauripus*) *E. hitchcocki*, *Eubrontes* (*Grallator*) sp., and the probable protosuchid crocodylomorph track *Batrachopus* sp. (Nadon & Middleton 47).

Station 6.4: Channel-fill and lacustrine sandstones with fish bone- and coprolite-bearing intraformational conglomerate at base. The intraformational conglomerate is apparently the lateral equivalent of or derived from the fish-bearing, thin-bedded sandstone exposed in the cliff face to the east. The only identifiable remains consist of scales and a ceratohyal of a large *Semionotus* sp.

Station 6.5: Poorly-exposed lacustrine strata of McCoy Brook Formation, with sand-patch cycles and gypsum nodules. These units resemble the Blomidon Formation, which they were previously confused with (e.g., Powers, 1916; Klein, 1962).

Turning around and looking southwest towards the islands it is clear that at this stratigraphic level we are quite high in the McCoy Brook Formation. The fault block we are standing on is in structural and stratigraphic continuity with the dip slope of the islands, particularly Moose Island, rather than the basalt ridge and Old Wife Point to the southeast. Assuming a rough estimate of 10° average dip gives us a stratigraphic distance of about 120 m from Station 6.5 to the top of the North Mountain Basalt, which may be stratigraphically higher than any of the other outcrops in the basin.

The next to the last island to the west is Pinnacle Island. On the south side of the island is a truncated and coarse-grained Blomidon Formation with the aforementioned red, gray, and black Partridge Island member. The Whitewater Member seems unusually coarse and thin here, although part of this thinning may be apparent because several faults are present. The Red Head member is also thin and seems to lack obvious eolian strata. However, the base of the member, and base of the Blomidon Formation, is well exposed in the tidal flat and accessible at low tide and rests with a profound unconformity 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 (Fig. 48). 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.

Return to vehicles (about 40 min to Stop 2.4).

Return to Hwy 2	4.0 km
Turn left at Hwy 2	26.2 km
Turn left at Main St	0.3 km
Turn left at 2 Island Rd	0.6 km
Turn right into parking area for Fundy Geological Museum, Park	0.1 km

Figure 48. View at low tide on the south side of Pinnacle Island looking north. In the foreground are east-dipping sandstones and conglomerates of the Wolfville Formation. In the background are north-dipping Red Head member sandstones of the Blomidon Formation, Between and to the right are the basal-most Red Head beds resting with a profound angular unconformity upon the Wolfville Formation.

Stop 2.4: Fundy Geological Museum

45° 23.965'N 64° 19.392'W

Main Point:

Collection of vertebrates, including dinosaurs from McCoy Brook Formation mostly of Wasson Bluff.

Nova Scotia's 24th museum the Fundy Geological Museum was 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 this author 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. Bones were recovered from ancient river channels, sand dunes and basalt talus slopes. Among these were skulls and jaws of very rare trithelodonts, a group closely related to mammals. (excerpted from http://museum.gov.ns.ca/fgm/).

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.

Return to vehicles and drive back to Parrsboro center 1.0 km

Field Day 3: Hanging Wall basement, Chignetgo Subbasin and Hanging wall of the Minas Subbasin

The depositional stratigraphy of Triassic-Jurassic deposits were dramatically modified adjacent to the strike slip Minas fault zone. The stratigraphy of the hanging wall strata of the Fundy basin is, not unexpectedly much simpler. On day 3 (Fig. 27) we will look at hanging wall basement of both the Chignetco subbasin (Stop 3.1) and the main basin (Stop 3.4) which in both cases contains significant prerift source rock and look at the locally basalmost synrift deposits which consist wholly of TS II, Wolfville Formation (Stops 3.2 and 3.3).

About 1:07 hr from Parrsboro to Joggins (Stop 3.1)

Leave from Parssboro center going N on Hwy 2 towards 12.0 km
Turn left at Boars Back Rd (unpaved) 13.2 km
Continue on Barronsfield Rd 6.8 km
Turn left at 242/Fundy Shore Scenic Dr/Glooscap Trail
Continue to follow 242/Glooscap Trai 5.9 km
Turn left to Joggins Fossil Institute

Stop 3.1: Joggins World Heritage Site

Main Points:

- Hanging wall basement
- Potential gas prone pre-rift source rock
- Famous fossil locality

Pennsylvanian age continental strata of the Cumberland basin (Fig. 49) outcrop extensively along the eastern shore of Chignetco Bay. Seismic reflection profiles reveal a thick synrift section of the Fundy basin below this bay in a portion of the Fundy rift called the Chignetgo subbasin (Withjack et al., 1995). At Joggins hanging wall basement in this area consist of coal-bearing strata of the Joggins Formation that may have produced gas where buried below synrift strata. The following summary of the Joggins site is paraphrased from Calder et al. (2005).

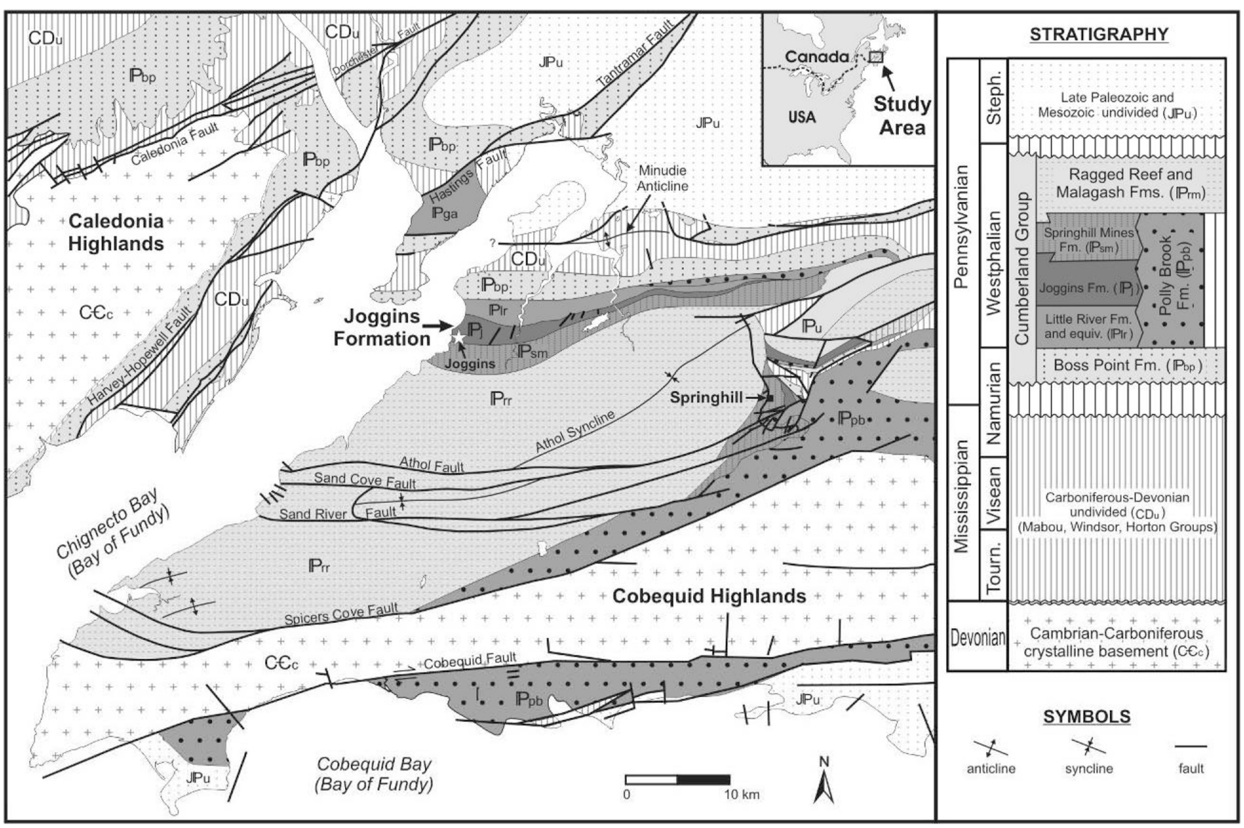


Figure 49: Map of the western Cumberland Basin from Calder et al. (2005). Stop 3.1 is at the arrow labeled Joggins Formation.

The cliffs at Joggins of Nova Scotia are amongst world's best exposure of coal-bearing Pennsylvanian strata. Exposed for several kilometers in cliffs the section gained initial fame largely through the publications of Sir Charles Lyell and Sir William Dawson beginning in the 1840s, and later through Darwin's On the Origin of Species, where the cliffs and its fossils figured prominently. To Lyell and Dawson, it clear that the Joggins strata preserved ancient landscapes with multiple horizons of standing trees, layers rich in charcoal from wildfires, and the world's earliest known amniote tetrapods ("reptiles") and land snails are notable. However, recent discoveries continue to be made and these include the recognition of restricted-marine faunal assemblages, and cycles that have been interpreted as sequences modulated by glacialeustacy. The wealth of historic and current research at this classic section resulted on July 7, 2008 in it being declared a UNESCO World Natural Heritage Site. For this visit we will arrive at high tide, so an extensive examination of the cliffs themselves will not be possible. We will visit The Joggins Fossil Centre where exhibits depict the fossil riches of the site.

The development of the Joggins succession may have been enhanced by syndeposition halokinesis of Windsor evaporites. According to Waldron & Rygel (2005), seismic profiles that the Joggins formation is preferentially thickened in the Athol syncline that itself is above a salt weld and that the Joggins Formation thins onto the flanks of adjacent evaporite-cored anticlines. Salt tectonics may have been equally important in the development of the early Mesozoic synrift strata as well.

Return to vehicles (about 3:10 hr to Stop 3.2)

Head northeast on Hurley Rd toward Main St	0.2 km
Turn right at Main St	0.2 km
Continue on 242/Fundy Shore Scenic Dr/Glooscap Trail	
Continue to follow 242/Glooscap Trail	9.6 km
Slight right at 242/Fundy Shore Scenic Dr/Glooscap Trail	8.7 km
Turn left at 302/Fundy Shore Scenic Dr/Glooscap Trail	8.8 km
Turn left at Hwy 2	2.2 km
Take the Transcanada Hwy NS/104 E ramp to Springhill/Truro/Halifax	0.7 km
Merge onto 104 E Partial toll road	96.7 km
Take exit 15 to merge onto 102 S toward Halifax/Truro	16.3 km
Take exit 12 for Collector HWY 289 toward Trunk HWY 2	0.3 km
Turn left at Pleasant Valley Rd	12.5 km
Slight left at Hwy 236	3.0 km
Turn right at Hwy 215	44.7 km
Park (with permission), walk north along path to beach, paralleling Rennie Brook	300 m

Stop 3.1: Hanging Wall Onlap Near Tenecape

45° 15.878'N, 63° 56.128'W

Main Points:

Hanging wall basement
Onlap of locally basal synrift strata
Fluvial sequence of lower Wolfville Formation

Excellent outcrops of the basal unconformity between the fluvial Evangeline Member of the Wolfville Formation in TS II and the pre-rift lacustrine Mississippian Horton Group occur at the small cove at the mouth of Rennie Brook, between East Walton and Tenecape. East Hants, Hants County, Nova Scotia (Fig. 50).

As we walk along the path down to the beach note the shale and sandstone outcropping along the way, These are pre-rift mostly lacustrine Mississippian Horton Group containing abundant fossils, in this area, especially tetrapod footprints and plants. The strata are strongly folded and most beds appear vertical although they are largely the limbs of folds. Just at the mouth of Rennie brook, there is the nose of a tight fold. Red and brown fluvial conglomerate, sandstone and bioturbated mudstone onlap the nearly vertical Horton strata.

Figure 50: Onlap angular unconformity between vertical Horton Group shales and sandstones and gently dipping Triassic Wolfville clastics at the mouth of Rennie Brook, stop 3.2. Note the similarity at this outcrop scale to the onlap relationship seen at the seismic scale in Fig.51.

Note that the Horton beds appear red, but close examination shows that structurally deep levels of the Horton are gray and black and the reddening appears to track the contact with the Wolfville Formation. A closer look shows that much red color is confined to joints and bedding planes and the cores of especially refractory units tend to be more pale or even grey. These observations suggest to us that the red color is due to Triassic weathering and is not primary.

The contact itself shows no hint of a residual paleosol other than the reddening of the Horton. The overlying Wolfville has abundant clasts of the Horton in the immediate vicinity of the contact (10s of centimeters), but further away from the contact there are few clasts of Horton

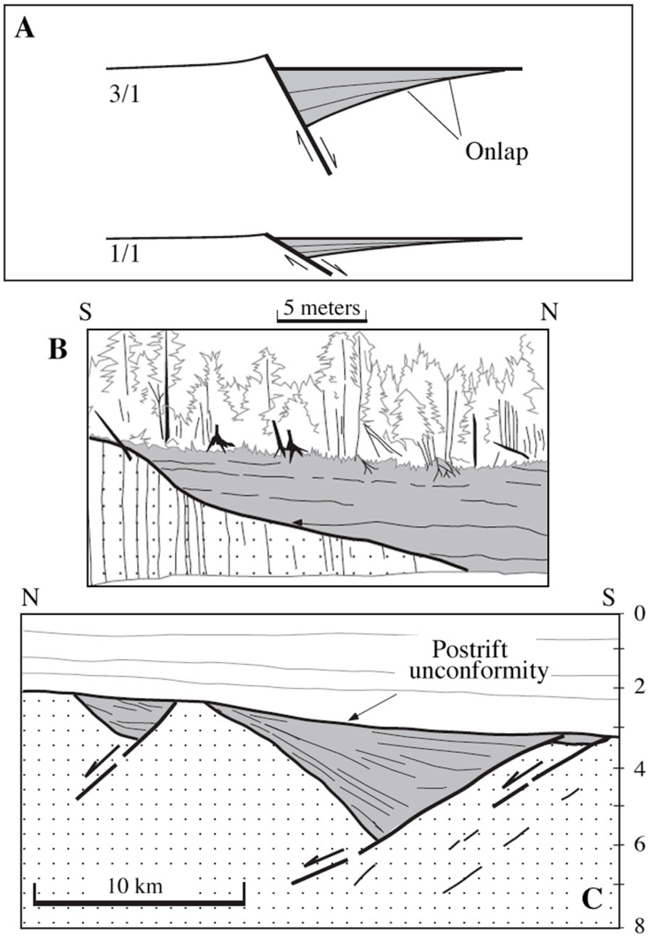
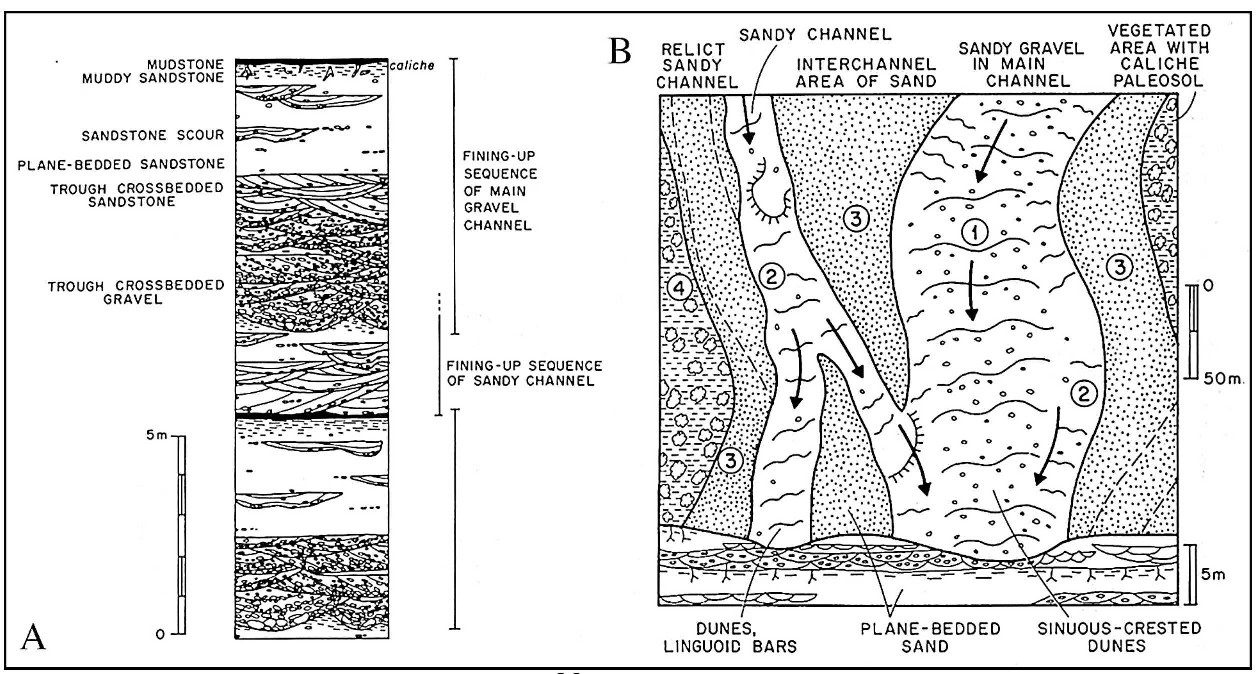


Figure 51: (Left) Hanging wall onlap patterns (modified from Withjack et al., 2002): A, transverse section (with and without vertical exaggeration) of a diagrammatic half graben – the magnitude of the footwall uplift and hanging wall subsidence decreases away from the border fault and sedimentary units thicken towards the fault and progressively onlap the prerift units; B, diagram of outcrop-scale onlap at Stop 2.2; C, line drawing of a seismic section from the Atlantis rift basin, offshore eastern North America (after Hutchinson et al., 1986; Schliche & Olsen, 1990).

Figure 52: (Below) A, Generalized fluvial fining-up cycles for the lower Wolfville at Tennecape based on Markov analysis. B, Model of plan view and cross section of pebbly-sandy braided river that produced the cycles in A. Arrows indicate direction of river flow. A and B from Hubert & Florenza (1988)



to be seen and the conglomerates evidently are not locally derived. It therefore appears as though the fluvial system stripped the soil from Horton flowing axially, an observation supported by the local paleocurrents that flowed to the northeast (Hubert & Florenza, 1988) strongly oblique to the current strike of the contact.

The bounding surfaces of the larger sediment packages within the Wolfville (e.g., dark line nearly half way up the right side of Fig. 50) clearly dip at a lower angle than the basal contact itself, the difference being a couple of degrees. Thus, the meter-scale bedding features progressively onlap the unconformity with the Horton Group. Similar features can be seen at other basal contacts along this shore (e.g. Fensom et al., 2006). This is exactly the relationship seen at seismic scale in many half graben where basal strata onlap basement, and is seen in the offshore eastern North America as well as other basins of various age (Fig. 51) (Withjack et al., 1995; 2002). This style of onlap is a consequence of the progressive tilting of the hanging wall during sedimentation that also produced thickening or fanning of strata towards the border fault and is a characteristic growth feature.

Lower Wolfville themselves consist of upward fining cycles of conglomerate, often clast supported, grading upward though a series of beds into pebbly sandstone, sandstone and sand mudstone. Most units are intensely bioturbated by invertebrate burrows and roots, although these are obvious only in the finer grained beds. The conglomerates and pebbly sandstones are usually trough cross-bedded. Hubert & Floreza (1988) show using Markov analysis of a broadly correlative section at Tenecape that the sequences were produced by a braided river system with gravel-sand bedloads and produced cycles 4-10 meter in thickness (Fig. 52).

No age-relevant fossils have been found in this area; however, this particular style of fluvial facies in the Wolfville tends not to produce bones. However, the facies relationship seen here in conjunction with the upward transition in facies from this type to on that does produce bones at near-by Tenecape described by Hubert & Forlenza (1988) suggest that these strata are older than the Carnian age assemblages known from the middle Evangeline Beach Member of the Wolfville Formation as seen at Stop 3.3 (Burntcoat Head). There is no evidence, however, that they are as old as the Economy Member which appears to be Middle Triassic in age.

Return to vehicles on Hwy 215 (15 min to Stop 3.3).

Proceed back northeast on 215/Glooscap Trail/Shore Rd toward	
Donald St Continue to follow 215/Glooscap Trail	11.3 km
Turn Left onto Burntcoat Road	2.9 km
Turn left to Burntcoat Lighthouse	0.6 km
Park, and proceed to beach. And proceed to the east along the outcrops.	

Stop 3.3: Middle Wolfville Formation at Burntcoat Head

45° 18.692'N, 63° 48.464'W

Main Points:

- Fluvial strata of the Wolfville Fm. Of TS II
- Potential reservoir facies with lateral continuity
- Rich tetrapod assemblage of Carnian age

Outcrops around Burntcoat Head reveal well over 150 m of the middle Evangeline Member of the Wolfville Formation (Fig. 53). This section resembles that of the upper parts of the Tenecape section describe by Hubert & Florenza (1988) and is most simply interpreted as being stratigraphically above that at Stop 3.2.

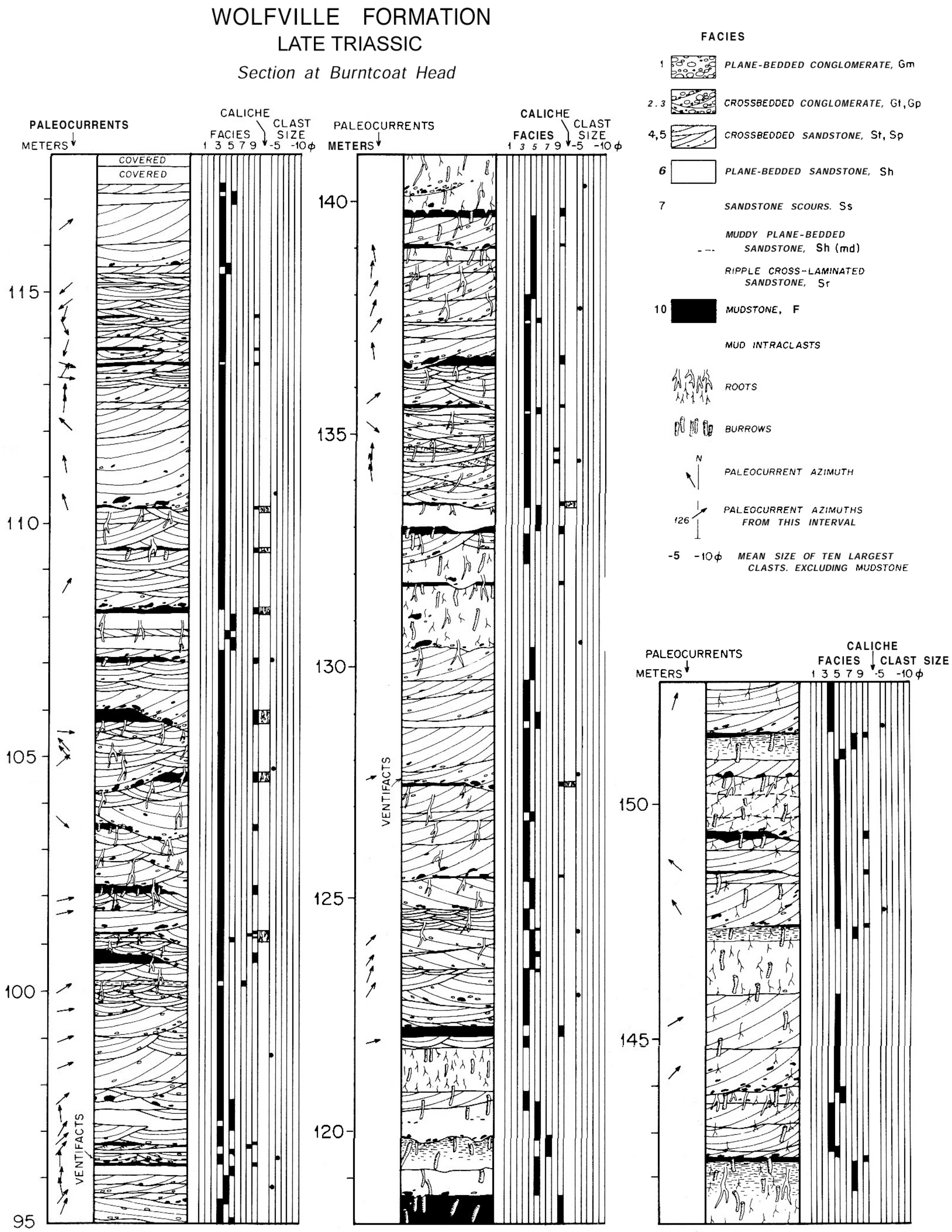


Figure 53: Section at Burntcoat Head (Stop 3.3) very slightly modified from Hubert & Florenza (1988) shoring Sedimentary facies and paleocurrents.

Overall, the section at Burntcoat differs from that at Stop 3.2 in being finer grained and more mud rich. With very little conglomerate with extra basinal clasts. Intraformational conglomerates are present, however, and these tend to be clast supported with calcite cement filling much of the interstices. According to Markov analysis by Hubert & Florenza (1988), the Burntcoat Head section also consists of fluvial cycles, here averaging 1.6 m in thickness, ranging from 0.5 to 4.5 m (Figs.53, 54). Their modal cycle (Fig. 54) begins with an erosional surface overlain by thick (> 30 cm) sets of cross-bedded sandstone where the lower sets contain mudstone intraclasts, often matrix supported with a calcite cement, with a few pebbles. Trough cross bed sets are about three times as abundant as planar sets thinning and fining upwards with most cycles ending with mudstones that contains root casts, *Scoyenia* burrows, and calcite nodules. Leleu et al. (2007) have suggested that the calcimorph paleosols and the major erosion surfaces can be correlated across the basin and similar kinds of sequences have the potential to demarcate sand-fairways for migration, in this potential reservoir facies.

Hubert & Florenza (1988) argue that the fining-up cycles were deposited by a braided river system based on the apparent low suspended load implied by the high ratio of sandstone to mudstone and the thin, discontinuous nature of the beds of mudstone. Virtually no sets of lateral accretion surfaces are present suggesting a lack of meander beds. Furthermore, ripple cross-lamination rare and climbing ripple cross-lamination appears absent which not be the case for point bars. Hubert & Florenza (1988) further argue that the rivers were wide and shallow based on the fact the cycles are thin. The cross-bedded sandstones can also be traced for more than 100 m along the present marine terraces and about 75% of the cross-bed sets are trough-shaped which suggests that trains of sinuous crested dunes.

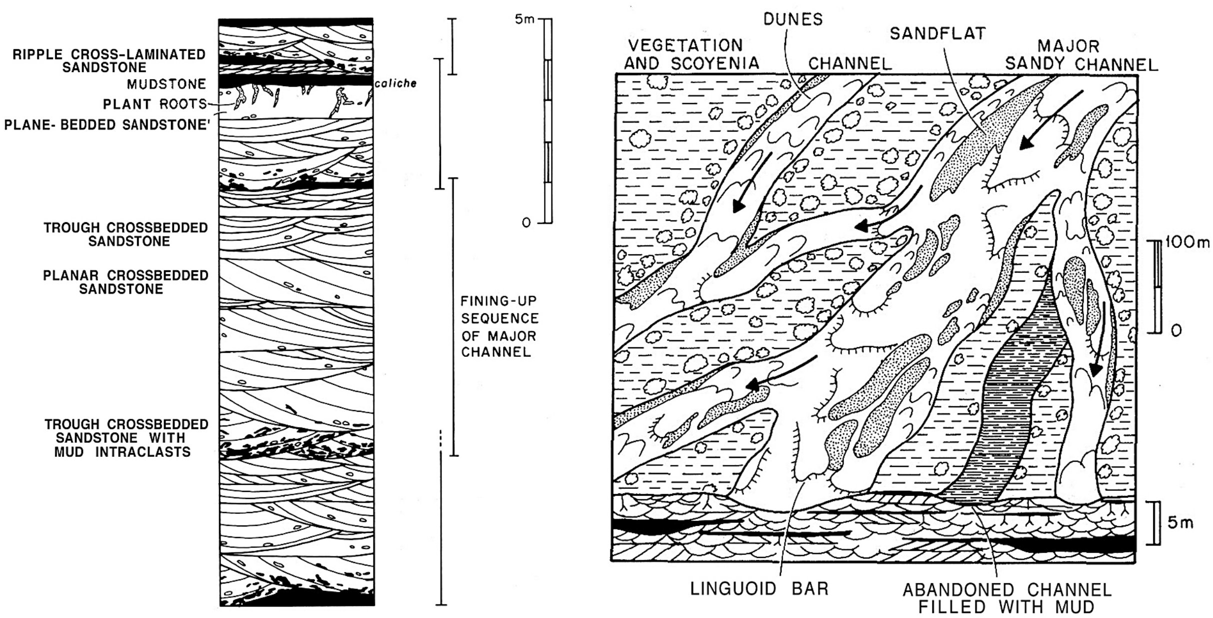


Figure 54: Left, Generalized fluvial fining upward cycles of sand-bedload rivers that built the 95-152-m interval at Burntcoat Head (Stop 3.3). Right, reconstruction of an active tract of the sand-bedload river that deposited the fining-up cycles shown on the left The river is sketched at moderate flow stage when dissected compound sandflats are emergent Both from Hubert & Florenza (1988).

The depositional environment of these fluvial systems has been interpreted as semi-arid because of the presence of paleosol caliche horizons, alluvial-fan conglomerates, and the presence of ventifacts. However, the caliches are not well developed and occur in very heavily

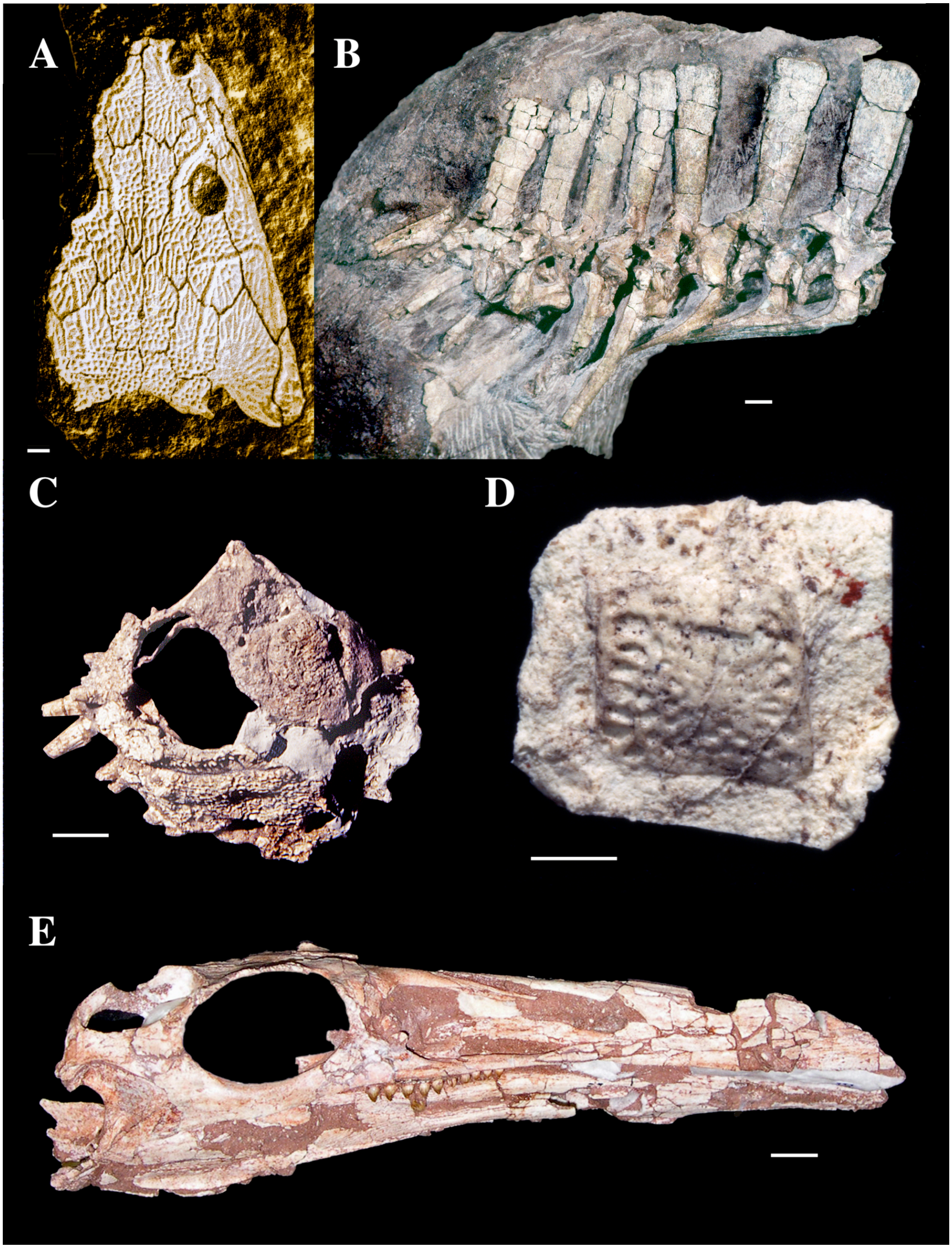


Figure 55: Examples of tetrapods from the Burntcoat Head area, middle Wolfville, Stop 3.3: A, the amphibian *Metoposaurus bakeri*; B, rauisuchimorph neck vertebrae; C, the armored procolophonid *Scoloparia glyphanodon*; D, osteoderm of the aetosaur ?*Stagonolepis*; E, the beaked trilophosaurid *Teraterpeton hrynewichorum*. Scale is 1 cm.

bioturbaed (heavily vegetated) units, and the ventifacts were clearly transported from their sites of origin and could have been much older than their deposition age (recycled from TS-I?). In contrast, perennial rivers are suggested by the presence of large (fist-sized) clams in conjunction with semiaquatic tetrapods notably large amphibians, in agreement wit the interpretations of Leleu et al. (2007). The paleosols are heavily rooted, suggesting significantly dense vegetation, and herbivorous tetrapods are the most abundant forms present. No eolian units are present, however, the semiarid interpretation of Hubert & Florenza (1988) was heavily influenced by concept that the Wolfville did contain significant eolian beds, but these strata have proved to be over 10 million years younger and in a different tectonostratigraphic sequence (i.e., Stop 2.3, Red Mead Member of the Blomidon Formation. In summary, compared to the fluvial systems of TS I and TS III, those of the middle Wolfville Formation were deposited in a significantly more humid environment, consistent with its pale geographic position closer to the equator (Fig. 9).

Tetrapod remains were first found by William Take in the 1950s and then were collected systematically by Donald Baird (Baird & Take, 1959; Baird, 1972) and colleagues during the 1960s though early 1980s. More recently significant material has been collected by avocational paleontologists, notably George Hrynewich of Dartmouth, as well as professionals notably Hans-Dieter Sues. The tetrapod bones and partial skeletons are fairly common at Burntcoat Head and comprise the best current evidence of its age. Fragmentary bones are most common in the calcite-cemented Intraformational conglomerate that also produces clams. However, partially to completely articulated remains have also been found in fine sandstones, and very recently associated material has been found in mudstones. The fauna has a lot of unique elements (Fig. 55), but shares some conspicuous taxa with other areas, perhaps most importantly *Metoposaurus bakeri*, otherwise known from the oldest assemblage of the Chinle Group of Texas recognized as the Otischalkian (Lucas, 1998). The presence of this taxon suggests via a long-distance global correlation web that the age of this middle Wolfville assemblage is Carnian, which is in agreement with the palynological age of the presumably correlative Echo Cove Member (Stop 1.3)

Amongst the unique taxa are the bizarre *Teraterpeton hrynewichorum*, which appears to be a Trilophosaurid, allied to the Otischalkian *Trilophosaurus*. However, the assemblage as a whole is unusual not only in the large number of seemingly endemic taxa, but especially in apparently lacking the semi-aquatic phytosaurs, which are the most common Late Triassic faunal element in the Triassic of the northern hemisphere.

Table 1. Vertebrates from the Wolfville Formation at Burntcoat

- Semionotidae
  - *Semionotus* sp.
- Temnospondyli
- *Metoposaurus bakeri*
- Traversodontidae
  - *Arctotraversodon plemmyridon*
- Dicynodontia indet.
- Procolophonia
  - *Acadiella psalidodon*
  - *Haligonia bolodon*
  - *Scoloparia glyphanodon*
- Sphenodontia indet.
- Rhynchosauridae
  - *Scaphonyx*? sp.
- Trilophosauridae
  - *Teraterpeton hrynewichorum*
  - Trilophosauridae indet.
- Rauisuchiformes indet.
- Crocodylomorpha indet.
- Aetosauria
  - ?*Stagonolepis* sp.
- Suchia incertae sedis
  - *Revueltosaurus* sp.

The most common animals are instead the procolophonids represented by three endemic forms *Acadiella psalidodon*, *Haligonia bolodon*, and *Scoloparia glyphanodon*. Carnian (sensu Muttoni et al., 2004) assemblages from closer to the paleoequator, such as those from the Richmond and Deep River basin have small traversodontids such as *Boreogomphodon*, probable ecological vicars of the procolophonids (Whiteside et al., 2006)as the most common faunal elements suggesting strong paleolatitudinally controlled faunal provinces. The only traversodontids from the Wolfville know thus far is the bear-sized *Arctotraversodon plemmyridon*. A list is given in Table 2. While phytosaurs are lacking other cruotarsians present include large unidentified rauisuchiformes, aetosaurs, and small, unidentified crocodylomorphs. Conspicuous by their absence are dinosaurs, remains attributed to dinosaurs in the literature from the Wolfville pertaining instead to either indeterminate archosaurs or the odd suchian group containing *Revueltosaurus* sp. (Irmis et al., 2007). Diagnostic dinosaurs are absent from other confirmed Carnian age deposits as well.

In the Argana and HABC basins the equivalent of the middle Wolfville is most simply hypothesized to be the Irohalen member of the Timesgadiwine Formation and the Oukaimeden Sandstones, respectively. However, the outcrops of the Evangeline Member seen here and at Stops 3.1 and 4.5 appear to be much further up along the hanging wall (and thinner) than their Moroccan equivalents.

Return to vehicles (About 1:42 hr to Stop 4.4).

Return to 215 /Glooscap Trail/Shore Rd	3.5 km
Go straight onto Singer Rd	5.0 km
Turn right at Noel - Gore Rd	9.4 km
Turn right at Hwy 236	0.3 km
Turn right to stay on Hwy 236	32.2 km
Slight left at 23	56 m
Slight left at 215/Glooscap Trail Continue to follow Glooscap Trail	8.2 km
Continue on Hwy 14	2.1 km
Turn right to merge onto 101	16.4 km
Turn right at Ben Jackson Connector	0.6 km
Turn left at 1	0.9 km
Turn right at Hutchinson Rd	1.5 km
Turn right at Blue Beach Rd	0.6 km

Stop 3.4: Horton Bluff Formation, Blue Beach (optional)

45° 6.049'N, 64° 12.917'W

Main Points:

Mississippian lacustrine potential source rock in hanging wall
Rich fossil assemblages including early amphibians

A significant extensional phase following the mid-Devonian Acadian Orogeny resulted in the formation of several half graben on and offshore in the Maritimes that filled with Horton Group deposits (Hamblin and Rust, 1989). The Horton is the oldest Paleozoic unit to overlap the boundary of the Meguma and Avalon terranes that docked during the Acadian orogeny. Grey and dark gray strata of the Late Devonian-Mississippian (-Tournaisian) Blue Beach and Hurd Creek members of the Horton Bluff Formation, of the Horton Group, outcrop at this stop at Blue Beach and Horton Bluff (Fig. 56). Here the strata exhibit no signs of paleoweathering.

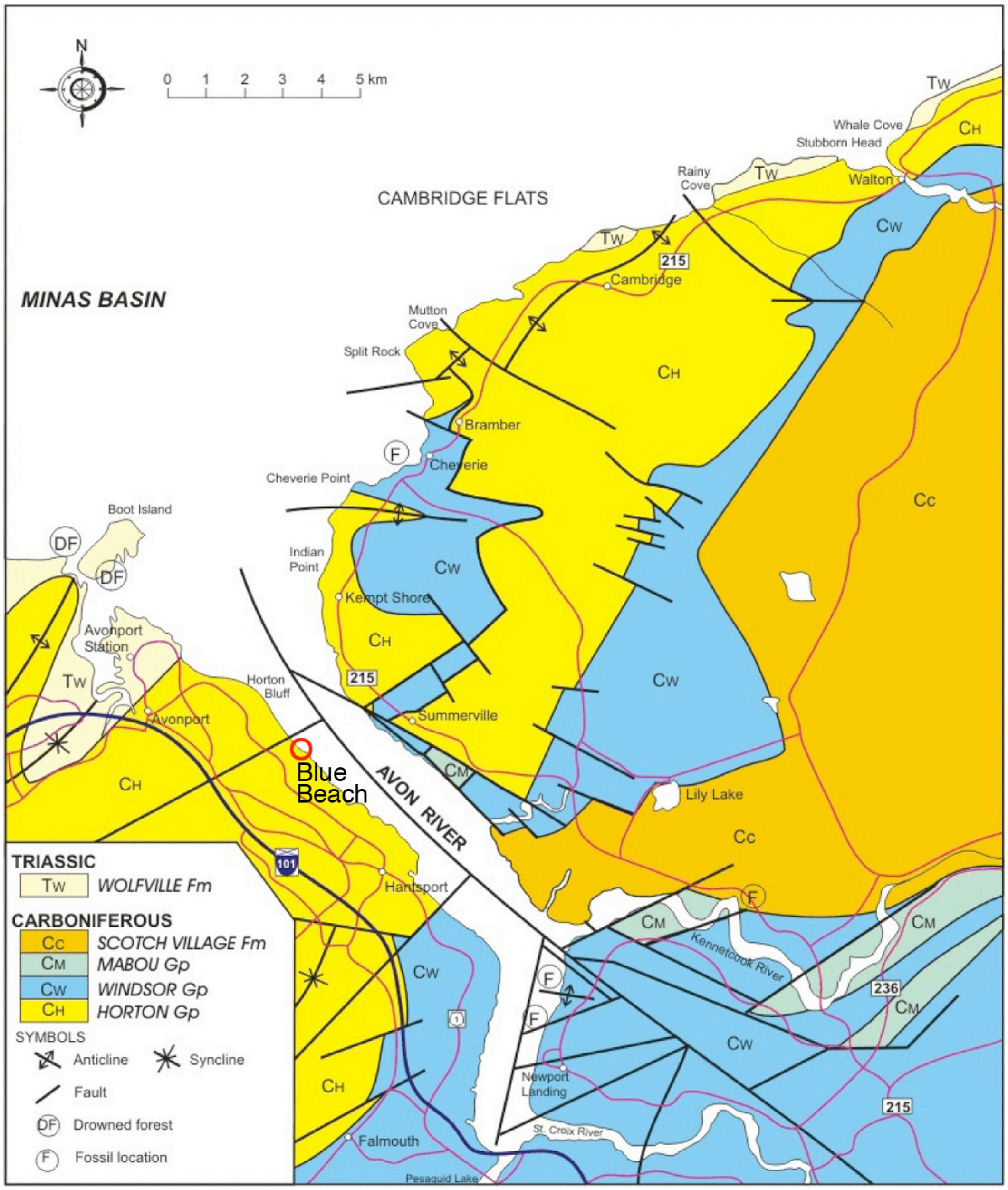


Figure 56: Map of the region around Blue Brook, Stop 3.4. From Skilliter et al. (2004).

One of the most prominent aspects of the cliff section, especially to the north is the well-developed cyclicity (Hesse & Reading, 1978; Martel & Gibling, 1996). The cycles shallow upward, consisting of basal grey shale; middle sandstone and shale; and an upper green mudstone with dolomitic beds and nodules and tend to be about 6 m thick. To out knowledge they have not be studied quantitatively.

In many ways, what Joggins is to the Pennsylvanian, Blue Beach is to the early Mississippian, with the area producing rich assemblages of plants, invertebrates, fishes, and amphibians. The footprint assemblage from here is arguably the oldest terrestrial assemblage in the world (Lucas et al., 2004). While the depositional environment has long been characterized as lacustrine, marine ostracodes and agglutinated foraminifera are present (Tibert, 1996; Tibert & Scott, 1999) suggesting at least some minimal connection to the sea (Calder, 1998).

Relevant to the goals of this trip is the fact that strata such as these underlay much of at least the northern Fundy basin and provide yet another albeit gas prone source rock. The organic matter in these rocks reached their relatively high thermal maturity in the Carboniferous (Mukhopadhyay, 1991). According to Calder et al. (1998) TOCs at Blue Beach rage from <1 to 3.5% but are thermally over-mature, with T_{max} values of 436-460°C and yields of <5 l/t on distillation, but only a few kilometers away at Falmouth the shale's are of substantially better grade containing 2.7-31.6% TOC, with hydrocarbon yields of up to 28.3 l/t and Tmax values of 433-439° C. There is considerable spatial heterogeneity in maturity and quality in these rocks. Caldor et al. (1989) also argue that there was a vascular plant rather than algal origin for most of the organic matter based on the low hydrogen indices (81-180) with kerogen types 2 and 3 being represented. Mukhopadhyay (1991) shows a relationship of maturity and distance form the Cobequid fault, part of the Minas fault system, with R0 values exceeding 1.5% near the fault and falling into the oil widow 2-3 km away from it.

The maturity profile of Horton Group source rocks indicates that they have been rendered over mature by fluid migration adjacent the Cobequid Fault where Ro values exceed 1.5%. At distances 2 to 3 km away from the fault, however, they fall within the oil window (Mukhopadhyay, 1991). Any additional effects of Triassic burial or later thermal events would most likely act to remobilize hydrocarbons already generated.

Return to vehicles (17 min to Old Orchard Inn).

Head northwest on Blue Beach Rd toward Bluff Rd	0.6 km
Turn right at Bluff Rd go 3.8 km	
Slight left at Oak Island Rd go 0.2 km	
At the traffic circle, take the 1st exit onto the 101 W ramp	0.3 km
Merge onto 101 W go 11.0 km	
Take exit 11 toward Trunk HWY 1/Collector HWY 358/Greenwich/Port Williams/Canning go	0.4 km
Turn left at Greenwich Connector	62 m
Continue on Greenwich Rd S go	0.2 km
Turn right into Old Orchard Inn parking area	0.2 km
Park	

Field Day 4: Hinge Margin of the Minas Subbasin

The stratigraphy from Wolfville, north to Scots Bay is the simplest we will see on this field trip. We will see the canonical section of the Fundy basin and see it in the order t would be revealed as if drilling down from the surface. Consequently Stops 4.1-4.5 proceed down section from the Scots Bay Member of the McCoy Brook Formation down through the North Mountain Basalt, Blomidon Formation, and Wolfville Formation. In most ways the stratigraphy is comparable to what we saw on Day 2, but it lacks the structural complications and instead reflects straightforward deposition on the hinge margin of the basin.

Leave for Stop 4.1 (about 32 min to stop)

Head northeast on Greenwich Rd S	0.2 km
Continue on Greenwich Connector	0.9 km
Continue on 358	10.3 km
Slight right at 221/358	0.1 km
Turn left at Chapel Rd	1.3 km
Turn left at Hwy 358	3.9 km
Turn left at Gospel Woods Rd	0.2 km
Turn right at Ross Creek Rd	6.2 km
Park and proceed toward Beach, ford Ross Creek and Head northeast along coast	

Stop 4.1: Scots Bay Member of the McCoy Brook Formation and Underlying North Mountain Basalt, Scots Bay area, Kings County, Nova Scotia

N45° 14.981' W064° 26.833' to N45° 15.087' W064° 26.643'

Main Points:

Basin-wide lacustrine limestone sequence on top of North Mountain Basalt
Extreme similarity to Moroccan basalt sequences

The base of the uppermost formation in the Fundy basin, the Scots Bay Member of the McCoy Brook Formation, consists of a cyclical sequence of white lacustrine limestones and gray, greenish, and red calcareous strata (Fig. 57). The sedimentary rocks of these sequences can be divided into six categories, each with characteristic non-marine fossils, as follows: 1) brown quartz-rich sandstone with dinosaur footprints; 2) thick-bedded, sometimes cross-bedded chert and limestone with clastic textures; 3) crudely-bedded, commonly silicified limestones with large desiccation cracks and few fossils; 4) stromatolitic limestones and cherts with domed heads and stromatolite tufas around trees; 5) red, green, and white crudely to thin-bedded snail and clam coquinas with probable charophyte debris; and 6) laminated green-gray calcareous siltstones with ostracodes, probable charophyte debris, and fish fragments, articulated remains and coprolites.

At these localities the Scots Bay member consists of three large-scale cycles 1-4 m) (Fig. 57B) (Suchecki et al., 1988), each consisting of a lower laminated, green, calcareous siltstone overlain by massive limestone and chert with abundant desiccation cracks. Each large cycle is composed of several shoaling-upward sequences marked by desiccation-cracked horizons suggestive of the smallest (decimeter-scale cycles) within the laminated clay-rich portions of the Blomidon Formation (Stop 4.3).

This member, seen here at its type area, appears at virtually all places the upper contact with the basalt is seen. It appears to be the correlative of the calcareous Durham Member we saw in the Silver Ridge B-1 core of the Shuttle Meadow Formation (Stop 2.3). A very similar

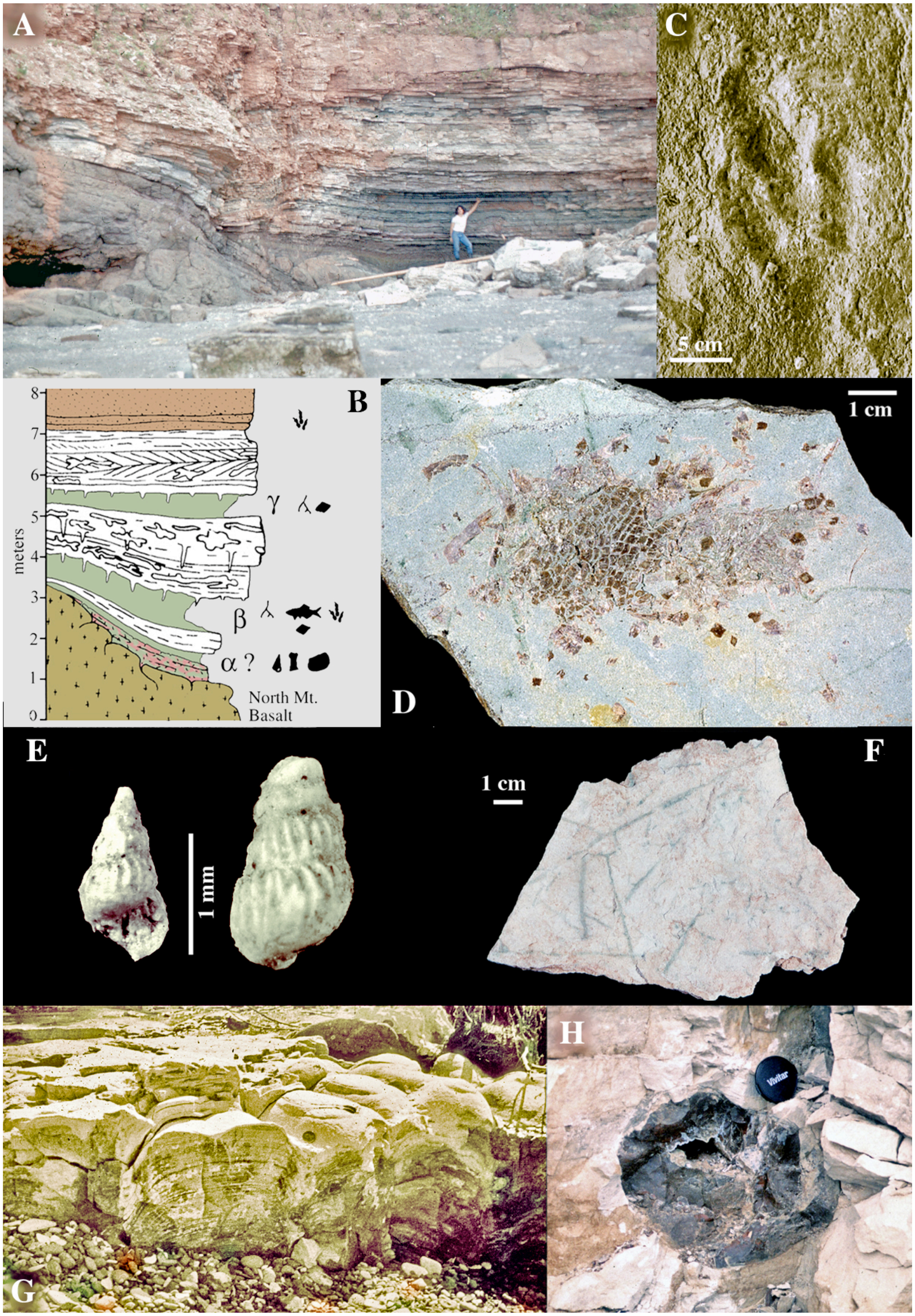


Figure 57: Scots Bay Member of the McCoy Brook Formation. A, thickest section of Scots Bay Mb. at station 3, Stop 4.1. B, Measured section at station 3, (α, β, γ refer to three deeper lake units). C, *Eubrontes* (*Anchisauripus*) *tuberosus*. D, *Semionotus sp*. E, hydrobiid gastropods. F, plant fragments. G, domal stromatolites. H, silicified stromatolites surrounding a tree branch or trunk.

limestone sequence occurs interbedded with the basalt over the western suite of the Moroccan basins, where it contains a very similar fauna, including some of the same species of mollusks. Suchecki et al. (1988) and Birney DeWet & Hubert (1989) argue on the basis of the presence of specific forms of chert and relatively heavy values of $\delta^{18}O$ that the Scots Bay Member lakes were fed by hydrothermal springs. This is possible, however given the basin-wide presence of these limestones, it seems simpler to explain the heavy $\delta^{18}O$ values as contribution of the buck rock chemistry by a later deep-ground water, possibly derived from the underlying basalt itself.

Despite the fact that the sequence is lacustrine, and is predominately non-red it has very low organic contents, in all cases lower than 1%. The rarity of articulated fish suggests, although fish are abundant, together with the absence of microlaminated units suggests that the water was never deep enough to chemically stratify, despite the fact the lakes were at least sometimes very large in area. Based on the abundant stromatolites and charophyte remains suggests high productivity, hence, ecosystem efficiency was very high and virtually all organic material produced was respired. The extraordinarily widespread occurrence of lacustrine limestones in this time interval suggests a very unusual climate system and but one still responsive to Milankovitch forcing.

Time permitting we will walk northeast to three areas along the upper contact of the basalt. Station 1 is at N45° 14.981' W064° 26.833' where a low series of outcrops reveals limestones with abundant stromatolites of two kinds: abundant stromatolites encrusting tree branches and trunks, and large domal stromatolites (Fig. 57G). The encrusting stromatolites have largely been replaced by chalcedony, and the trees nearly completely removed leaving a natural mold of their exterior, which is often filled with quartz and chalcedony with geopedal geometries. It appears as though the trees were drowned by the encroaching lake with erosion exposing the bases of the trees and shallow root systems, which subsequently were encased in stromatolites, the wood finally rotting out with some of the trees collapsing.

At station 2 at N45° 15.060' W064° 26.709' a several meter-thick brown sandstone that locally caps the limestones has partially collapsed revealing dinosaur footprints on its upper surface (Fig. 57C). This or a similar sandstone exhibits planer cross-stratification and seems to cap the carbonate section in this area.

The thickest section of Scots Bay Formation is at station 3 at N45° 15.087' W064° 26.643' (Fig.. 57A. About 8 m of section is preserved. This is the most fossiliferous of the sections, with relatively common articulated and disarticulated fish and rare sphenodontian bones in one bed (Fig. 57D) and silicified clams and snails common in another (Fig. 57E), present with charophyte debris and sparse crocodile (protosuchid) bones. *Semionotus* sp. is the only fish so far recognized (Olsen et al., 1982; 1989). However, large, *Semionotus*-containing coprolites suggest the presence of a large coelacanth or a shark as at Wasson Bluff (Stop 5.3) in the lateral equivalent of these beds.

Obvious at this outcrop is an onlapping of beds against the basalt at the NE and SW end of the small cove. While this might suggest that these are products of local ponds, the fact that the same basic stratigraphy reoccurs in other pods indicated that larger lakes were present, but that waves reworked units on highs into lows.

There is a striking similarity between the facies and stratigraphy of the Scots Bay Member in this area and what it seen in the Moroccan basins, particularly in the HABC. There, limestone sequences containing a virtually identical suite of facies and fossils occur interbedded with the basalts in a very specific stratigraphic sequence relative to the chemistry of the basalts

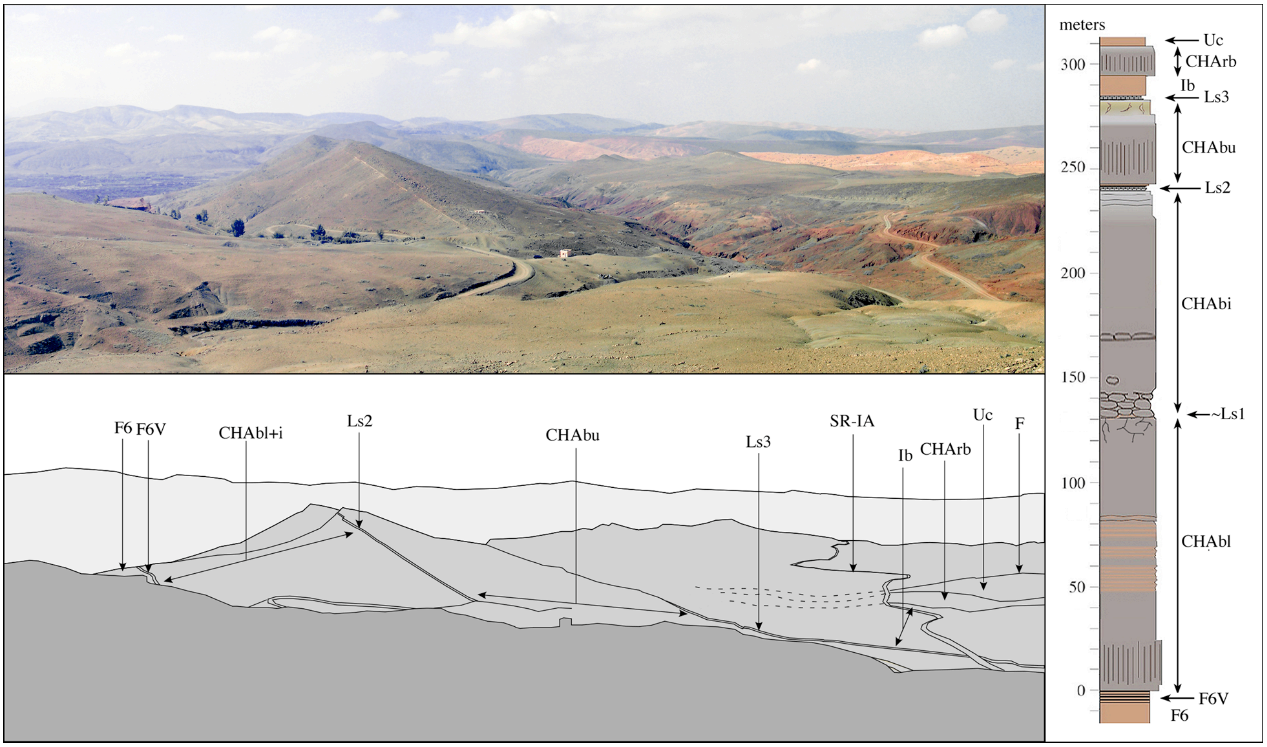


Figure 58: Panoramic view looking west at 31° 36.026' N and 007° 25.553' W (~6.5 km southeast of Sidi Rahal, Morocco) of Central High Atlas Basalt succession near locality producing hydrobiid gastropods. Note that the basalt units and limestone and clastic interbeds are in unquestionable, direct superposition in these outcrops, and show unequivocally that the hydrobiid gastropods come from units between the intermediate and upper CHA basalt units. Abbreviations are, from left to right: F6, Triassic age formation F6 of Mattis (1977); F6V, variegated, beds of uppermost F6, thought by us to contain the Triassic-Jurassic boundary; CHAbl+I, Central High Atlas lower and intermediate basalt units; Ls 1, approximate position of limestone interval at Tiourjdal, does not outcrop in this immediate area; Ls2, limestone and red bed interval with hydrobiid gastropods (actual locality is at 31° 36.365' N and 007° 26.029' W at the village of Ait Tamret)); CHAbu, Central High Atlas upper basalt unit; SR-IA, unpaved road, road between Sidi Rahal and the salina near Imiz Ar; Ls3, limestone unit directly above upper basalt unit; Ib, interbed of red clastics between upper basalt unit and recurrent basalt; CHArb, Central High Atlas recurrent basalt unit; Uc, Upper red unit with gypsum; F, fault (Whiteside et al., 2008).

(Fig. 58). The same kind of onlapping geometry typifies the equivalent limestones in the HABC, where the same very unusual species of hydrobiid snail is found (Whiteside et al., 2007, 2008) (Fig. 59).

Kontak (2008) recognizes three distinct mappable members within the North Mountain Basalt: East Ferry, Margaretsville and Brier Island members. According to his mapping of the members, the Scots Bay Member rests on the Margaretsville Member, the Brier Island Member not making it this area. If the Margaretsville Member is not simply a facies of the Brier Island Member, the map pattern can be interpreted and meaning that the Brier Island Member is younger than the exposed Margaretsville Member at this stop and could be younger than at least the lower part of the Scots Bay Member as seen here. At the present time, although considerable high-quality geochemical data is available for the North Mountain Basalt, the analyses have not been carried out stratigraphically, so we do not know if the different members might differ geochemically. As pointed out previously, however, it does seem as though there may be some geographic heterogeneity (e.g., Deneen, 2007). In any case, the chemistry of the North Mountain basalt does seem to overlap in some way with both the lower basalt and intermediate basalt formations of the HABC.

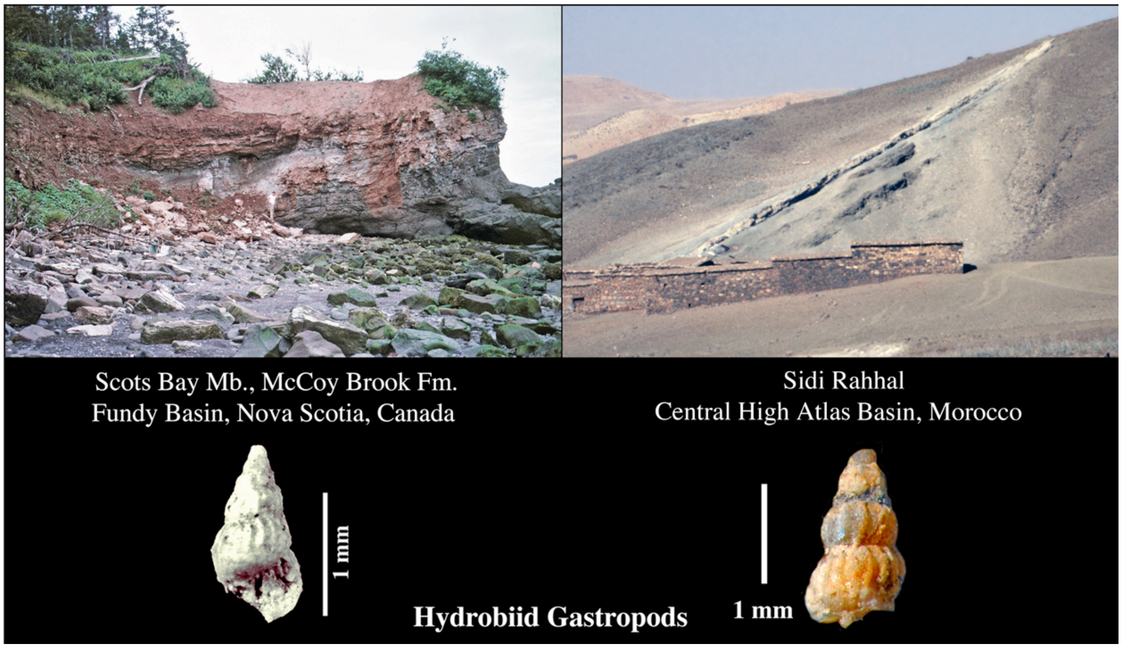


Figure 59: Limestone sequences of the Scots Bay Member of the McCoy Brook Formation of the Fundy basin, Nova Scotia (left) at Stop 4.1, and the limestone sequence between intermediate and upper basalt units in the Central High Atlas basin at Sidi Rahal, Morocco and examples of their dwarf hydrobiid gastropods (from Whiteside et al., 2007; 2008).

The classic paper of Carmichael & Palmer (1968) did treat the paleomagnetic sampling of the North Mountain Basalt stratigraphically. What is more, they found a significant difference, largely in declination between the lower two-thirds of the formation and the upper third (Figs. 60, 61). Interestingly, this difference is roughly the same as that found by Marzoli et al. (2004) and Knight et al. (2004) within the HABC basalts. In the case of the HABC basalts this change in declination occurs below a thin interval of reverse polarity that Marzoli et al. (2004) correlated (incorrectly in our view) with E23r, which lies below the Tr-J palynological transition. In as much as no similar declination transition occurs anywhere else in HABC basalts, and this shift presumably reflects secular variation, we tentatively conclude that the two shifts are correlative. Unfortunately, Marzoli et al. (2004) and Knight et al. (2004) show this transition in two different places relative to the geochemistry of the basalts. In Marzoli et al. (2004) the declination transition occurs within the intermediate chemistry basalts while Knight et al, show it between the lower and intermediate basalts (see Fig. 61). One tantalizing additional piece of evidence is that Carmichael & Palmer have one point (20 in Fig. 60) that could be of reverse polarity (although it could also be misoriented).

The point of this digression is that the paleomagnetic data would seem to indicate that the declination direction change should occur within the Margaretsville Member or between the Margaretsville Member and the Brier Island Member of the North Mountain Basalt and that the reverse polarity zone should be above this, plausibly either in the base of the Scots Bay Member or within the sedimentary interbed between the Margaretsville Member and the Brier Island Member as seen in Digby to Brier Island area (i.e., the AV-C-1-2 core - Westport), which could, based on the mapping of Kontak (2008), be one and the same thing. We (and D.V Kent) are in the process of testing this hypothesis by additional paleomagnetic sampling.

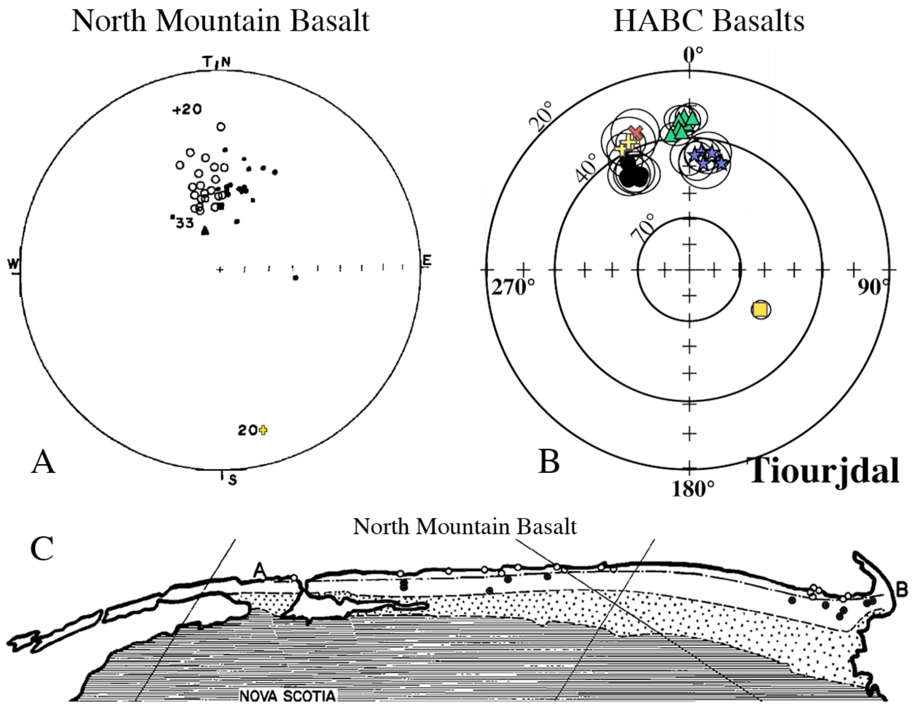


Figure 60: Comparison of tilt corrected paleomagnetic data from the North Mountain Basalt (Carmichael & Palmer, 1968) and the HABC basalts (Tiourjdal area) of Morocco (Knight et al., 2004). A, Sample directions of NRM of North Mountain basalt (from Carmichael & Palmer, 1968) using the conventions of the equal-area projection: TN is geographic north. Solid circles indicate, north-seeking directions downward, lower stratigraphic samples; open circles, north-seeking directions downward, upper stratigraphic samples; plus signs, north-seeking direction upward; square, axial dipole field; triangle, present field. Sample 20 that has a north-seeking direction upward direction that could be a reverse. B, Projection of mean flow directions from Tiourjdal, HABC, Morocco (from Knight et al, 2004). Symbols and colors correspond to “DGs” from Fig.61A identified by single-cluster analysis of mean flow directions. Note the similarity to A, with more scatter in A probably because the samples are not averaged by flow. C, Geological map from Carmichael & Palmer (1968) showing sampling sites in A. Line A-B separates stratigraphically lower samples to the southeast from higher stratigraphic samples to the northwest. We have colored the lower samples black to conform to the plot in A.

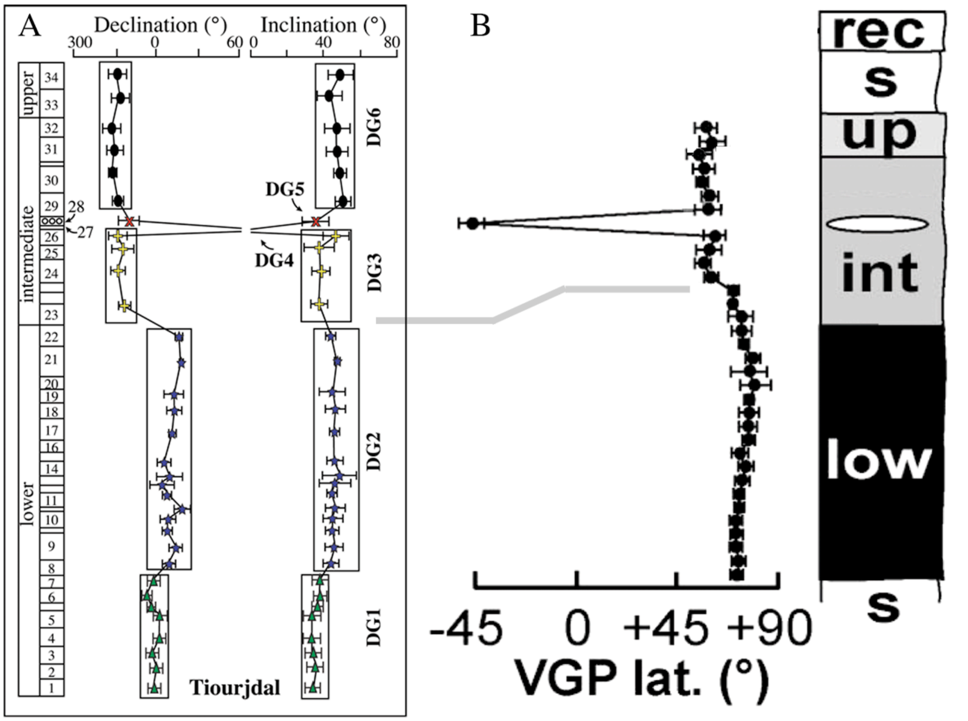


Figure 61: Comparison the same paleomagnetic data from Knight et al. (2004) and Marzoli et al. (2004) from the Tiourjdal section (HABC, Morocco). A, Magnetostratigraphy, mean flow declination, and inclination and directional groups (DGs) from Knight et al. (2004). B, Schematic stratigraphy and magnetostratigraphy (virtual geomagnetic pole [VGP] latitudes with A_{95} errors) for basalts (low, lower; int, intermediate; up, upper; rec, recurrent) of representative central High Atlas section (at Tiourjdal) from Marzoli et al. (2004). Gray line shows match in declination change in both versions, but not difference in the identity of which basalt chemistries that declination change falls in.

Because the lower basalt and lower intermediate basalt formations of the HABC of Morocco are the only plausible pre-extinction CAMP flows yet identified (as argued by Marzoli et al., 2004; Knight et al, 2004), the striking similarities with the North Mountain Basalt, where the boundary occurs BELOW the basalt (e.g. Stop 2.1), makes investigation of the latter's correlation to Morocco of paramount importance to understanding the cause of the Triassic-Jurassic extinction.

Return to vehicles (35 min to Stop 4.2)

Go back south on Ross Creek Rd	6.2 km
Turn left at Gospel Woods Rd	0.2 km
Slight left at Hwy 358	4.3 km
Turn right at Stewart Mountain Rd	3.1 km
Head northeast on Pereau Rd toward Mill Creek Rd	4.7 km
Park in parking area for beach	

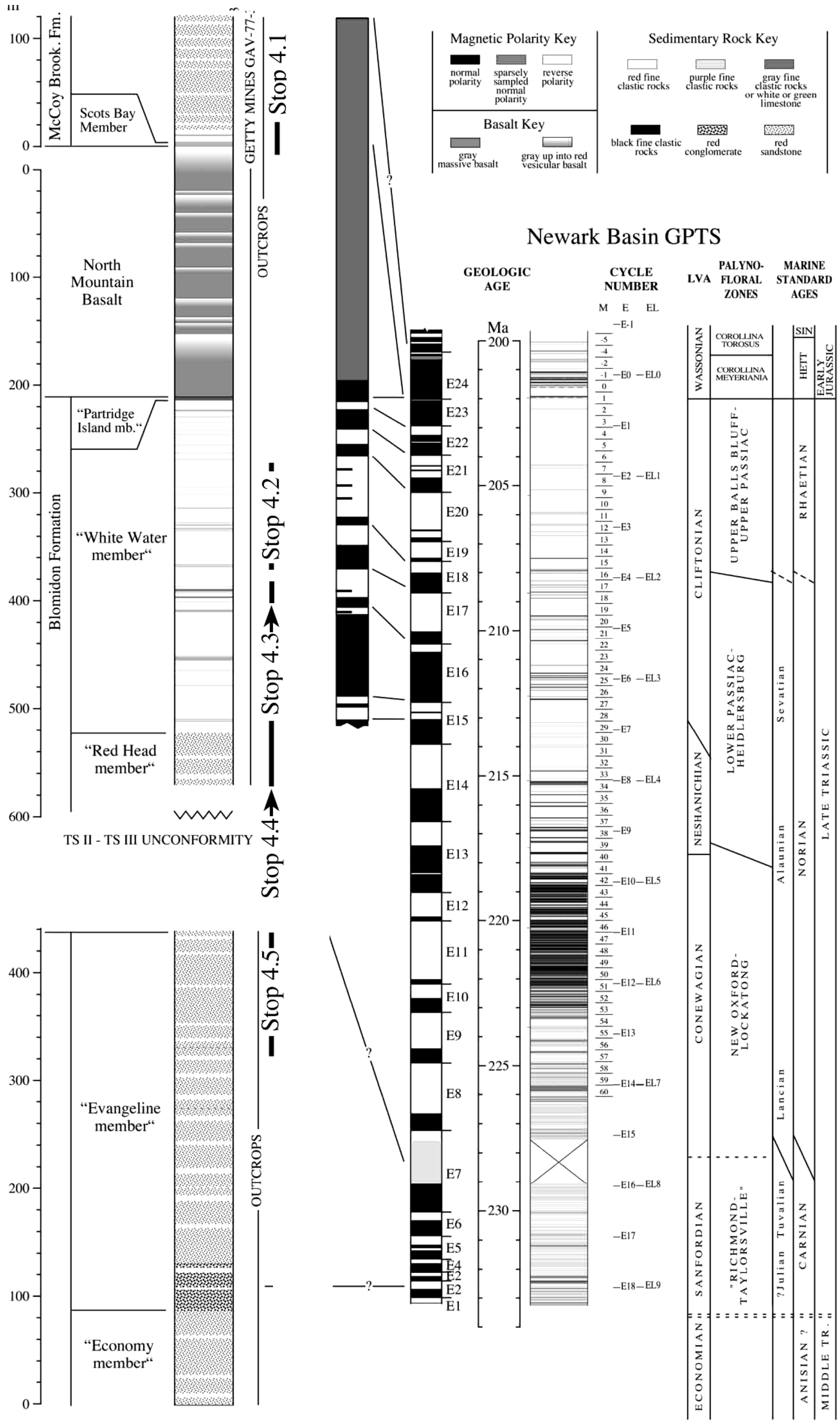


Figure 62: Fundy basin section correlated to the Newark basin astronomically calibrated geomagnetic polarity time scale showing the position of field stops 4.1 – 4.5. Based on Kent & Olsen (2000), Olsen et al. (2005a,b).

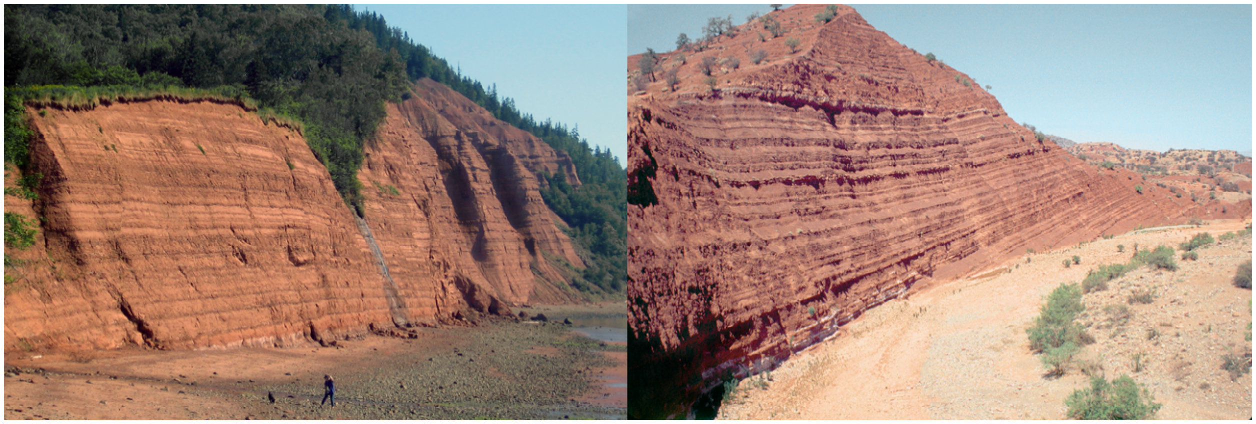


Figure 63: Comparison between Whitewater Member of the Blomidon (Stop 3.2), left, and Hassein Member of the Bigoudine Formation, right. Left photo by Erin Ostli in Google Earth (ID: 831228).

Stop 4.2: White Water Member of the Blomidon Formation, White Water, Kings County Nova Scotia

45° 15.344'N 64° 20.951'W

Main Points:

Sand patch cycles in huge cliffs
Cyclical hierarchy

This stop consists of a 2 km walk down section through about 100 m of middle Blomidon Formation, beginning at the White Water picnic area of the Blomidon Provincial Park (Fig. 62). The section of the Blomidon Formation is perhaps the best example of the profound similarity between the facies of the Moroccan basins and the Fundy basin (Fig. 63). The similarities exist all scales and suggest control by overriding climatic and tectonic processes.

Walk down the stairs leading to the beach at the east end of the picnic area and look north towards the spectacular cliff outcrops of the upper Blomidon Formation and lower part of the lower flow of the North Mountain Basalt. Most of the Blomidon Formation outcrops consist of cyclic red and brown lacustrine sequences producing almost no fossils. The uppermost Blomidon, however, consists of the Partridge Island Member, that looks almost exactly like it does Stop 2.1.

Turn south and proceed south, following cliffs. Distances are measured cumulatively southeastward from the foot of the stairs. Distances were originally measured in paces by Olsen et al. (1989) with one pace = 0.9448 m..

62 m: Borden Brook-well-developed sand-patch cycles. At these outcrops the sand-patch cycles average about 1.5 m thick. Sand-patch cycles can be generally traced at least the length of outcrops (300-1000 m).

Sand patch cycles are a variety of Van Houten cycle dominated by sand patch fabric described originally by Smoot & Olsen (1985, 1988) (Fig. 64). Sand-patch fabric consists of irregular pods of sandstone and siltstone with cuspate contacts with surrounding mudstone, internal zones of different grain sizes, and sometimes cross-laminae showing no consistent orientation relative to other pods. The sand-patch fabric may be cut by jagged-edged, mudstone-filled cracks. Sand patch fabrics have been identified with certainty in the Fundy and are widely distributed in the basins of Morocco (Smoot & Olsen, 1988; Olsen, 1997). Some beds in the lower Boonton Formation of the Newark basin also appear to have this fabric. Beds containing the sand-patch fabric are part of the playa sequence itself.

The sand-patch fabric commonly occurs in meter-scale cycles termed sand-patch cycles that make up larger cycles. Most sand-patch cycles (Smoot & Olsen, 1985, 1988) consist of a lower massive red mudstone with irregular wispy sandy pods succeeded by a red brown massive mudstone-sandstone showing a very well-developed sand-patch fabric. Above the unit with a sand-patch fabric is often a faintly ripple-bedded sandstone. Some cycles may begin with a laminated claystone or fine siltstone, in some places underlain by a thin oscillatory-rippled sandstone or siltstone. Irregular tabular zones at various intervals in the cycles are riddled with euhedral crystals or pods of gypsum, which produce vugs where weathered. Sand patch cycles are typical of the Fundy-style lacustrine sequences.

213 m: Very thin (20-30 cm) sand-patch cycles with well-developed claystone bases make up marker beds which can be traced to cliffs on north side of steps from the picnic area

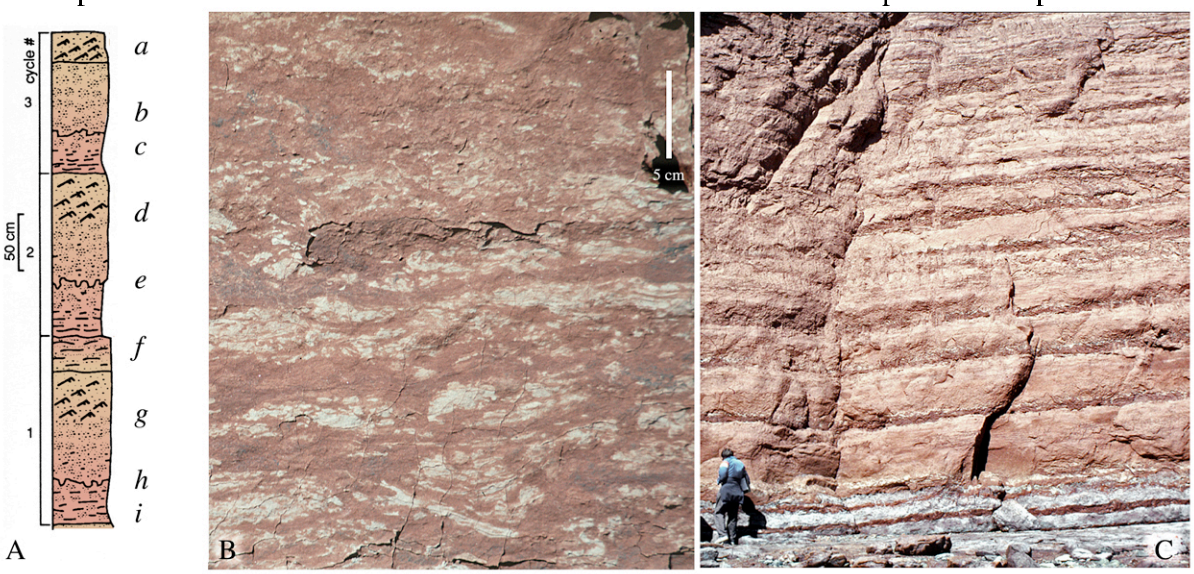


Figure 64: Sand patch cycles and fabrics. A, diagram of sand patch cycles: *a*, brown muddy sandstone having irregular ripple cross-laminated lenses; *b*, brown sandy mudstone having sand-patch fabric; *c*, red mudstone having irregular green mottles and wispy lenses of sandstone; *d*, brown sandy mudstone having sand-patch fabric grading up into muddy sandstone with irregular ripple; cross-laminated lenses; *e*, red mudstone having irregular green mottles and wispy lenses of sandstone; *f*, irregular, lenticular thin beds of brown sandstone and red mudstone defining deformed ripples – layer boundaries are indistinct; *g*, brown sandy mudstone having sand-patch fabric grading up into muddy sandstone with irregular ripple cross-laminated lenses; *h*, red mudstone having irregular green mottles and wispy lenses of sandstone; *i*, brown muddy sandstone having irregular wispy lenticular beds (deformed ripples?). B, Close up of sand patch fabric. C, Series of sand patch cycles at Stop 4.2.

allowing sections to be correlated unequivocally. Clusters of such thin sand-patch cycles often underlie clusters of three or four thicker (1-2 m) sand-patch cycles, making a higher-order cycle

499 m: Mudstone cycles consisting of about five 20-to- 40-cm-thick beds of laminated claystone and siltstone grading upward into massive mudstone, with sand-patch fabrics becoming common in the upper beds. Such cycles increase in frequency lower in the Blomidon

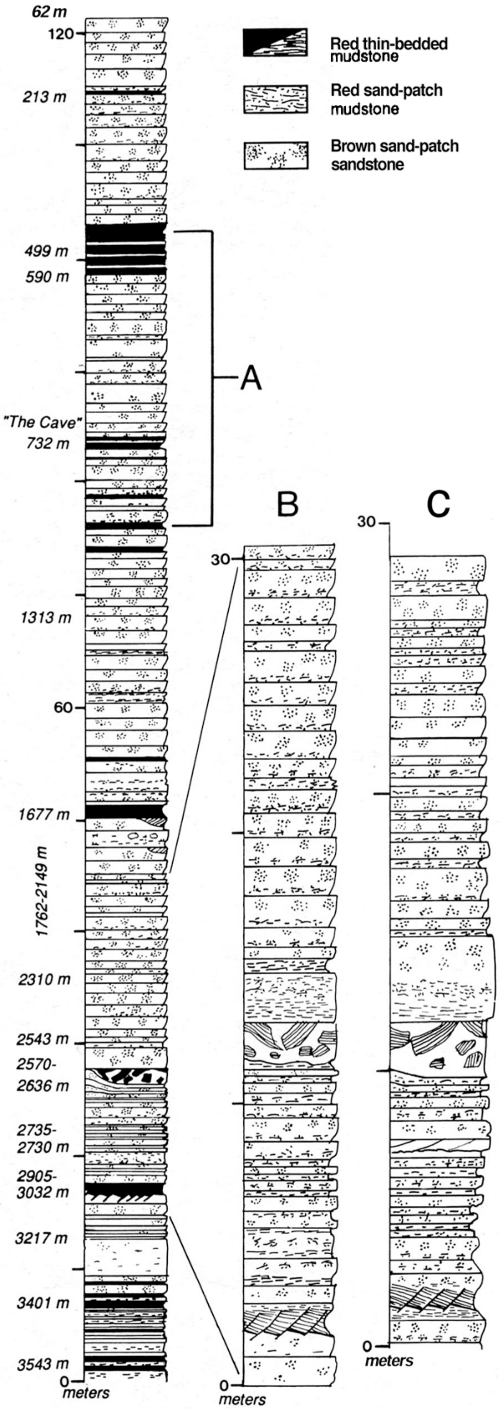


Figure 65: Section of the Blomidon Formation at Blomidon Provincial Park, Stops 4.2 and 4.3. A, Location of data used for spectrum in Figure 64: B, enlargement of section at meter-marks 1782-3032; C, same stratigraphic interval as (B) but repeated At Stop 4.3.

Formation, and these mudstones are the main source of fossils. Alternations of two or three of these cycles with two or three sand-patch cycles make up larger cycles similar to those made solely of sand-patch cycles (as at 144 m).

590 m: Long, thin sandstone lenses, possibly out in cliff face in lower beds of mudstone cycles. lenses pinch out down-dip to beach level.

732 m: Two well-developed mudstone overlain by a series of well-developed sand-patch making a yet higher-order cycle (Fig. 65). Fourier analysis section (Fig. 66) shows compound periods of 0.91 m, 1.24 m, 4.10 m, and the well-developed mudstone and sand patch cycles correspond 1.24 m thick cycles in the power spectrum. The from these mudstone rich lavers to the next cluster cliff (at 499 m) corresponds to the 4.1 m cycle. Tied to the GAV-77-3 core paleomagnetically, and hence to the Newark basin, Fourier analysis of the section seen at 732 m supports the meter-scale cycles are the ~20 ky Van Houten cycles, Fourier analysis also yields significant periods at around 100 ky, and 400 ky. However, without direct correlation to the Newark basin, the large amount of noise in this spectrum makes an argument based only Fourier analysis untenable, but that correlation, the orbital timing of the cyclicity is clear.

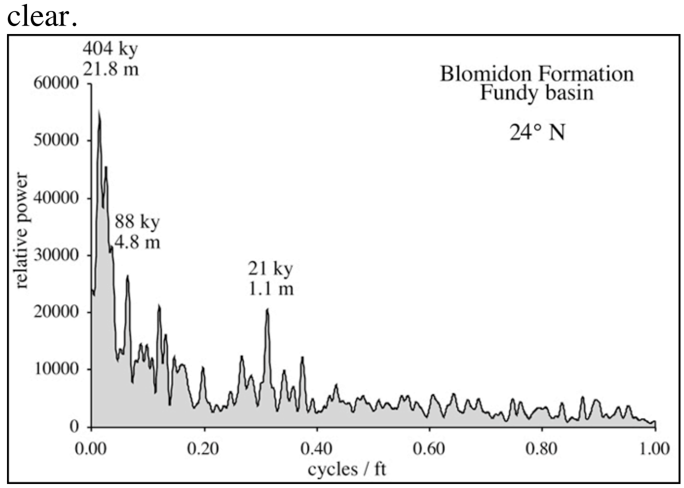


Figure 66: Power spectrum of section A in Fig.65A. The series of peaks centered around 1.1 m correspond to the sand patch cycles.

1162 m: Hanging outlet of Straddle Leg Brook. More sand-patch cycles.

1313 m: Very clear, well-developed sand-patch cycles.

1677 m: A well-developed mudstone cycle is on the west, with finely-laminated clam shrimp (bivalved branchiopod crustaceans) and ostracode-bearing claystone with pinch-and-swell siltstone laminae. On the south is a "reef" of eolian sandstone.

1708 m: Remnants of steel stairs from Glooscap Lane development.

1762-2149 m: Bowl-shaped deformation pods within well-developed sand-patch cycles are well displayed here.

Asymmetrical fanning, isolated beds of very well-sorted sand, and lack of associated or compensating uplift suggest progressive subsidence and filling of a very small basins, by dissolution of underlying evaporites.

Return to vehicles (10 min to Stop 4.3).

Head south on Pereau Rd toward Stewart Mountain Rd	4.7 km
Turn left (southeast) on Stewart Mountain Road	0.4 km
Park near beach	

Stop 4.3: White Water Member of the Blomidon Formation, Blomidon, Kings County Nova Scotia

45°13'4.37"N, 64°21'39.80"W

Main Points:

Large halite dissolution features

Walk north along beach to large outcrops on left.

First outcrops are red to greenish-gray, finely laminated claystone (90 cm thick) with thin pinch-and-swell siltstone laminae overlying a massive sand-patch-bearing sandstone and siltstone. This is the base of the section in Fig.65C. The unit is deeply and polygonally cracked and produces clam shrimp, darwinulid and other ostracodes, and semionotid and possibly redfieldiid fish scales. The unit apparently forms the base of a well-developed 2-m-scale mudstone cycle. The fossil-bearing unit is cut by east-trending listric and planar faults that rotate the units downward to the north. There is some evidence of rotation and syn-faulting deposition of eolian sandstone in one or two of the mini-half-graben. Although there is no distinct upper erosional surface, the faulting appears not to penetrate more than 4 m above the fossil-bearing unit, either soling out into roof faults or dying out as growth faults. Lower parts of faults appear to sole out into underlying sand-patch and gypsum- rich unit. Deformation is consistent with a volume loss in adjacent strata related to evaporite dissolution.

Further up section we see a sequence of thin sand-patch cycles about 8 m thick with a higher percentage of mud and gypsum than usual. Many low-angle faults are present.

This is followed upward by Brick red mudstone sequence 1.9 m thick with a flat bottom overlies a strongly lenticular (0-2 m) chaotic-looking unit consisting of decimeter- to meter-scale irregular pods and fissures of brown mudstone to coarse sandstone with internal bedding and abundant small (< 1 cm) to large (> 1 m) clasts of a distinctive deeply mudcracked, finely-laminated red, green-gray, purple, and yellow claystone containing fish scales, sparse articulated

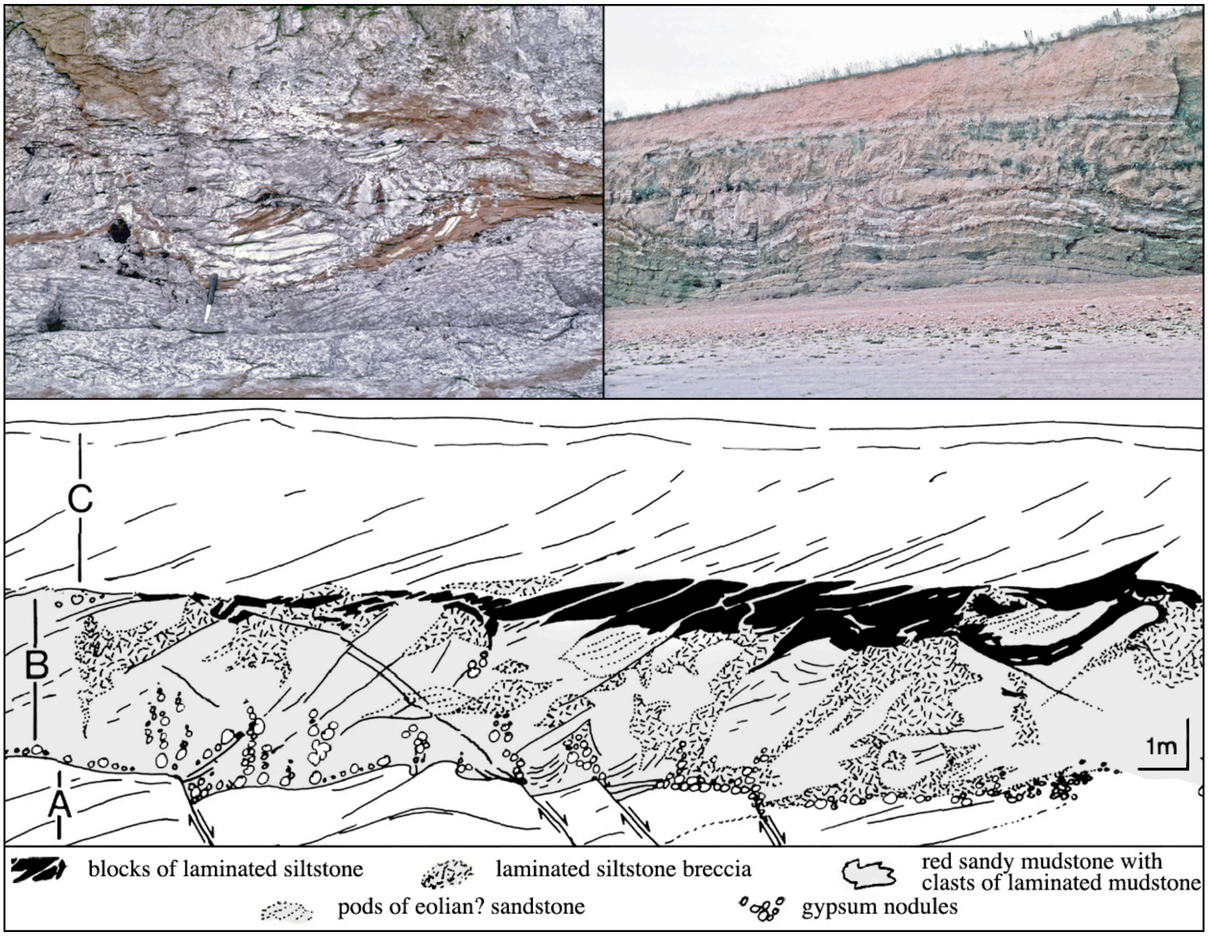


Figure 67: Salt dissolution structures at Stops 4.2 and 4.3: left, salt dissolution bowls with clear growth structures at 1762-2149 m (note hammer for scale) at Stop 4.2, right; large salt collapse structures at 2570-2636 m of Fig. 64 at the same stratigraphic level at those at Stop 4.3 (cliff is about 15 m high); below, one of the large salt collapse structures (above right); A) Underlying normal-faulted interval of sand-patch massive mudstone; B) dissolution interval; C) red mudstone interval. Note the downward projecting units of laminated shale breccia, apparently originating from collapse of the overlying units (modified from Olsen et al., 1989).

fish, and abundant clam shrimp and ostracodes (Fig. 67). Clasts seem to be derived from the bed that is occasionally present at the top of the lenses. The lower contacts of the lenses are conformable with underlying mudstone and sand-patch-bearing beds that are warped into gentle anticlines and synclines (or domes and bowls) that decrease in amplitude down-section. The whole section below the brick red unit is riddled with pods and euhedral crystals of gypsum. Pods and fissures in lenses containing clasts derived from overlying beds suggest progressive erosion and deposition below the sediment surface mostly prior to the deposition of the flat-bedded brick red mudstone. A meter-scale bed of salt could have been dissolved in stages and replaced by subsurface channels and fissures, or space for subsurface deposition could have been created by earthquake-induced surface ruptures, blowouts, and fissures. However, because no evidence of upward movement of material has been found, we favor the salt dissolution model. This stratigraphic sequence of solution collapse structures is repeated by faulting twice to the south. The southernmost, and hence most hingeward section at Delhaven has the collapse structures invaded by a fluvial sandstone that exhibits clear growth structures, related to evaporite dissolution. These evaporite dissolution structures have been described in detail by Ackermann et al. (1995), who conclude that the pods represent the dissolution of several meters of an evaporite mineral, probably halite. The finely-laminated red, green-gray, purple, and yellow claystone bed occur at 391 meters in the GAV-77-3 core (Fig. 64). The lake that produced these units was not sufficiently deep to avoid respiration of organic matter, even though it was deep enough to be occasionally below wave base.

The end of the outcrop consists of a long sequence of sand-patch cycles. At the other side of Mill Creek, the section repeats itself because of fault with normal apparent displacement.

These dissolution structures attest to the presence of bedded halite that was destroyed by subsequent penecontemporaneous dissolution similar to what is seen at the Triassic-Jurassic boundary sequence in the Argana basin (Fig. 29C, D).

Return to vehicles (10 min to Stop 4.4)

Go back to Pereau Rd,	0.4
Turn left heading southwest on Pereau Rd toward Mill Creek Rd	3.2 km
Turn left at Jackson Barkhouse Rd	0.9 km
Turn left at N Medford Rd	1.3 km
Drive forward along unpaved road to parking area	0.4 km

Stop 4.4: White Water and Red Head Members of the Blomidon Formation, "Paddy Island" area, Medford, Kings County Nova Scotia

45°13'4.37"N, 64°21'39.80"W

Main Points:

Base of TS II
Fluvial and eolian deposits
Faunal facies comparable to lower Bigoudine Fm., Argana Basin Morocco

The sea cliffs in the vicinity of Paddy Island, West Medford, Kings County, display the lower Blomidon Formation (Figs., 62, 68) including the basal member of the formation, the Red Head Member that we also saw at stop 2.3.

At the base of the small creek that leads down from parking area off North Medford Road, we can see the transition between the upper beds of the Red Head and overlying basal beds of the Whitewater Member of the Blomidon Formation. At this locality, 1-2 m intervals of brown, dune-scale, trough cross-bedded conglomerate and sandstone, passing upward into

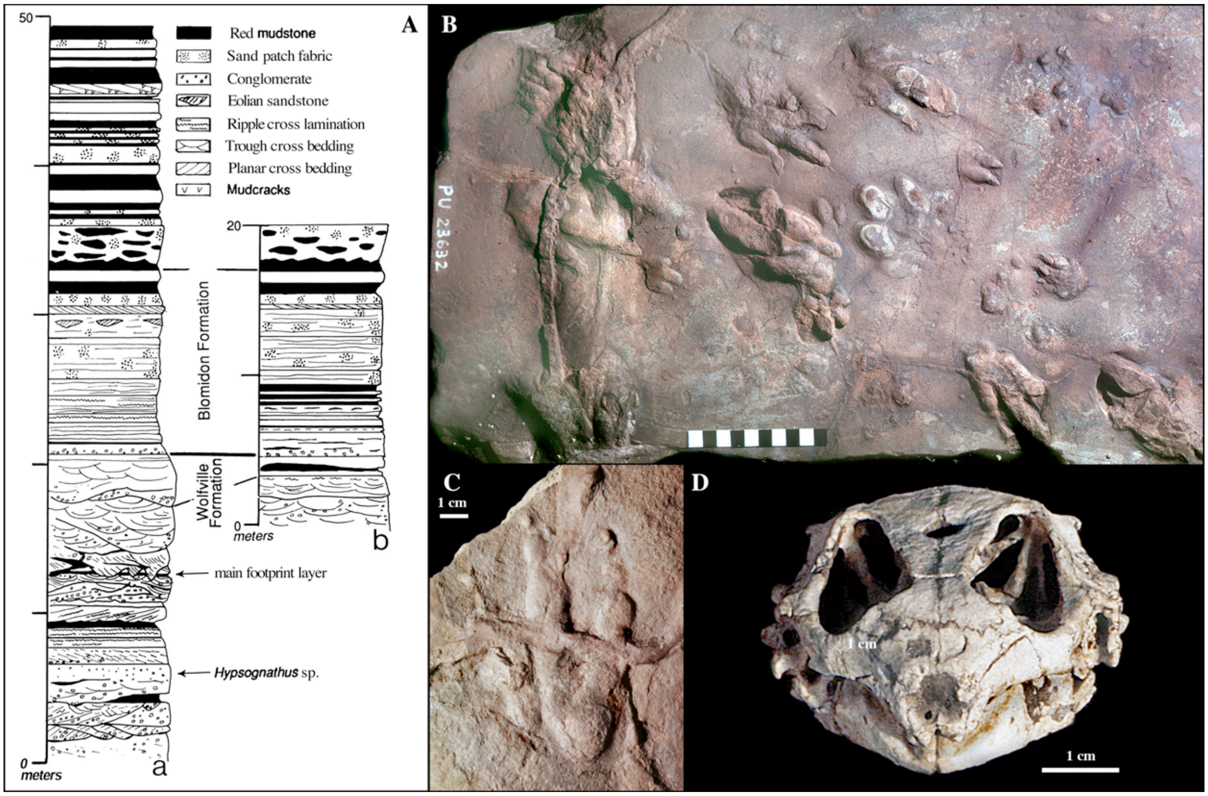


Figure 68: Paddy Island area, Stop 4.4: A, measured sections of the uppermost Wolfville Formation and lowermost Blomidon Formation adjacent to Paddy Island - (a) section on mainland south east of Paddy Island, (b) section covering the same interval as (a) but repeated by a normal fault and lying about 600 m to the north; B, Slab with several trackways of *Atreipus acadianus* and one of *Brachychirotherium* cf. *parvum*; C, new dinosaur-like ichnogenus; D, skull of *Hypsognathus* cf. *fenneri*.

smaller-scale, cross-bedded and ripple-bedded sandstone and planar-bedded red mudstone comprise 4- to 10-m-thick fining-upward cycles. The transition into the overlying member of the White Water Member of the Blomidon is marked by the disappearance of dune-scale trough cross-bedded units and conglomerate (Fig.68). The basal White Water beds are marked by planar, ripple-bedded, or massive, laterally continuous (at the outcrop scale) sandstone beds, and higher in the section by laminated to massive red claystones and mudstones and by massive sand-patch fabric. Sand-patch cycles occur at this outcrop, but are best seen higher in the Blomidon, at Stop 4.2. A higher order cyclicity can also be seen at these outcrops, consisting of intervals with well-developed sand patch cycles alternating with intervals of sandstone and mudstone without sand-patch fabric. The cycles average 6.6 m thick at these outcrops. These larger Blomidon cycles seem to pass, in phase, into the predominantly fluvial cycles of the uppermost Red Head Member of the Blomidon. The middle portions of the dune-scale cross-bedded conglomerate and pebbly sandstones appear to be the homotaxial equivalent of the sand-patch intervals, whereas the red, footprint-bearing mudstone intervals appear equivalent to the mudstone intervals in the Blomidon which mostly lack the sand-patch cycles. If true, this would imply a climatic control of the large-scale fluvial cycles, rather than an origin from the lateral migration of channel systems.

On the other hand, it is possible that the sand-patch fabrics have no counterpart at all in the fluvial cycles.

A fault cutting just east Paddy Island and int the cliff face repeats the basal Blomidon section and gives us a chance to see the lateral change in facies toward the basin interior (Fig. 68). As we might expect, the frequency of mudstone beds in the basal Blomidon increases basinward, and the upper portions of a conglomerate in the uppermost Red Head Member in the southern outcrops is replaced by a mudstone and sandstone sequence in the northern outcrops (Fig. 68).

Walk south from where we came down from the dirt road. The first promontory is comprised of sandstone and conglomerate of the upper Red Head member. Well-preserved reptile footprints are abundant in the laterally continuous claystone and siltstone beds in the uppermost at this locality (Fig.68). The most abundant taxon is the lizard like form cf. *Rhynchosauroides* sp., followed by numerous and well-preserved *Atreipus acadianus* (Olsen & Baird, 1986). The brachychirotheriid *Brachychirotherium* cf. *parvum* is also rather common. A new genus of track is also present but rare ("*Coelurosaurichnus* sp. 2" of Olsen, 1988b). In total, the assemblage is directly comparable to that seen in the lower to middle Passaic Formation of the Newark basin. A skull and disarticulated skeleton of the procolophonid *Hypsognathus* cf. *fenneri* was found at the 6 m position in poorly-sorted pebbly sandstone (Sues et al., 2000).

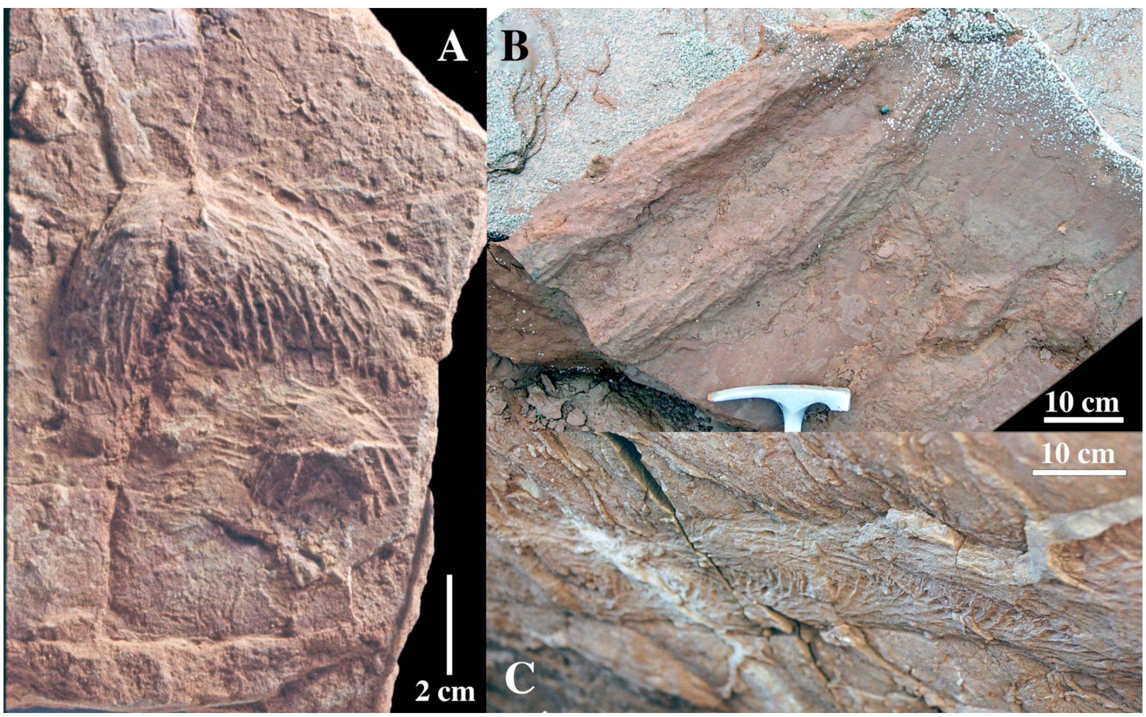


Figure 69: Tetrapod burrows from the Red Head Member of Stop 4.4 and the Tadrart Ouadou Member of the Argana basin, Morocco. A, Two terminations of burrows on underside of a sandstone over a claystone mainland cliff, Stop 4.4; B, Large burrow (left) and termination right, on underside of a sandstone over a claystone on the north side of Paddy Island. C, large burrow underside of sandstone bed, Tadrart Ouadou Member of the Bigoudine Formation, Argana basin, Morocco.

The footprint-bearing strata of the Red Head Member on the mainland coast and especially on the north side of Paddy Island proper have produced a remarkable assemblage of tetrapod burrows otherwise known from the Blomidon Formation at Rossway south of Digby, Nova Scotia, and the Tadrart Ouadou member of the Bigoudine Formation of the Argana basin (Fig. 69), that is homotaxial (but older) to the Red Head Member, and which contains significant eolian sandstone (see Stop 2.3, Fig. 45)

These burrows consist of a cylinder of sandstone or more rarely siltstone with bilateral sets of scrape marks on their upper and lower surface, the lower of which seem finer than the upper. Sometimes the bilateral sets of scrape marks meet in mid-line, sometimes they are separated by a smooth area. The largest widths are in excess of 18 cm and the smallest less than 5 cm. The burrows often track above a mudstone layer, and when they intersect a mudstone layer obliquely, the burrow terminates with a crescent set of bilateral scrapes.

That these are tetrapod burrows is attested to by the fact that the scrapes occur in sets of at least four corresponding to the four most prominent fingers of the primitive tetrapod manus. It is not obvious what kind of animal made these traces, but it could not have been an amphibian because they lack claws, which this burrow maker clearly had. A procolophonid such as *Hypsognathus* that constitute the most abundant skeletal remains from these units would be plausible candidate for the burrow maker.

In addition to being of interest in a biological sense, these burrows are useful sedimentologically. They unequivocally demonstrate that the strata in which they occur spent significant time after deposition above the water table, because non-amphibian tetrapods breather air and could not survive in a terminal, subaqueous burrow.

Walking further to the southeast we see a bed of probably eolian strata interbedded with fluvial sandstones and conglomerates. Below this is long section of fluvial gravels and sandstone in which root traces and invertebrate burrows are conspicuous by their absence. This is a major difference between the fluvial sequences of TS III and TS II in which root traces and burrows are very abundant (see Stop 4.5).

Return to Vehicles (10 min to 4.5)

Go back to Medford Rd	0.4 km
Head south on Medford Rd toward Weaver Rd	1.8 km
Turn left onto Clarke Rd and drive to end and park (ask Permission)	
Walk down path to beach	

Stop 4.5: Evangeline Member of the Wolfville Formation

45°10'39.97"N, 64°21'29.03"W

Main Points:

Upper TS II

Partly gray and purple fluvial and lacustrine Wolfville Formation

Correlative to lacustrine facies in Timesgadiwine Fm of Argana basin

The section of sea cliffs and tidal ridges at Medford Beach to Kingsport (Fig. 70) reveal fluvial and weakly developed lacustrine strata in the locally upper part of TS II, the Evangeline Member of the Wolfville Formation.

Walk south along the sea cliffs. There are abundant, well-developed red and purplish rooted calcic paleosols initially and then as we proceed south we will see yellow-weathering, gray conglomerates and then purple red and ostracode-bearing lacustrine mudstones in a facies much more humid looking than further up section (Fig. 70) in the Red Head Member of the Blomidon (TS III, Stop 4.4). These strata belong to TS II. Unfortunately, a covered interval to the north prevents us from seeing if there is evidence here for an unconformity between TS II and TS III.

The presence of marginal lacustrine conditions suggests there may be better-developed lacustrine facies deeper into the Fundy basin towards the depocenter that might resemble the cyclical lacustrine strata of the Irohalen Member of the Timesgadiwine Formation of the Argana basin (Fig. 8), but this has yet to be tested by drilling.

Return to vehicles (1:58 hr to Halifax)

Head west on Clarke Rd toward Medford Rd	0.2 km
Turn left at Medford Rd	0.2 km
Slight right at Parker Rd	1.3 km
Slight left at Jackson Barkhouse Rd	0.3 km
Turn right at Longspell Rd	0.3 km
Turn right to stay on Longspell Rd	0.3 km
Turn right at 221	4.2 km
221 turns slightly left and becomes Sheffield St (358)	1.5 km
Continue on 358	8.8 km
Continue on Greenwich Connector	0.9 km
Turn left onto the 101 E ramp to Windsor/Halifax	0.4 km
Merge onto 101	75.6 km
Take exit 1 on the left for Hwy 1	1.0 km
Merge onto 1/Bedford Hwy	16 m
Merge onto 102 S via the ramp to Halifax	17.4 km
Slight right at Bayers Rd	1.5 km
Continue on Young St	0.5 km
Turn right at Robie St	2.8 km
Turn right at University Ave	0.3 km
Make a U-turn at Seymour St	40 m
End of Field trip, Dalhousie Student Union, 6136 University Ave, Halifax, NS, Canada	

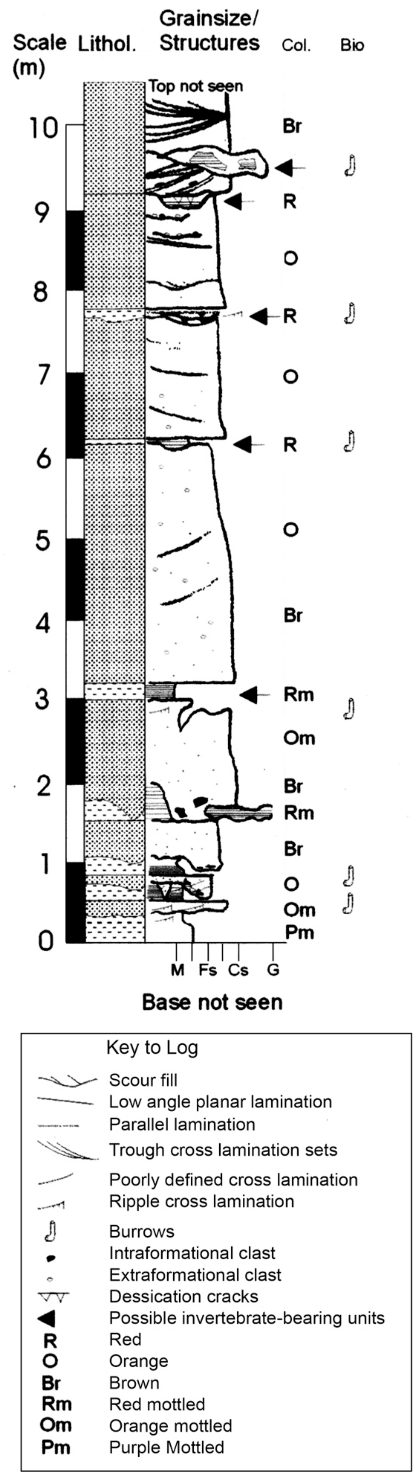


Figure 70: Section in Wolfville formation (TS II) at south end of transect of Stop 4.5 (courtesy of B. Cameron, 2001).

Acknowledgements

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