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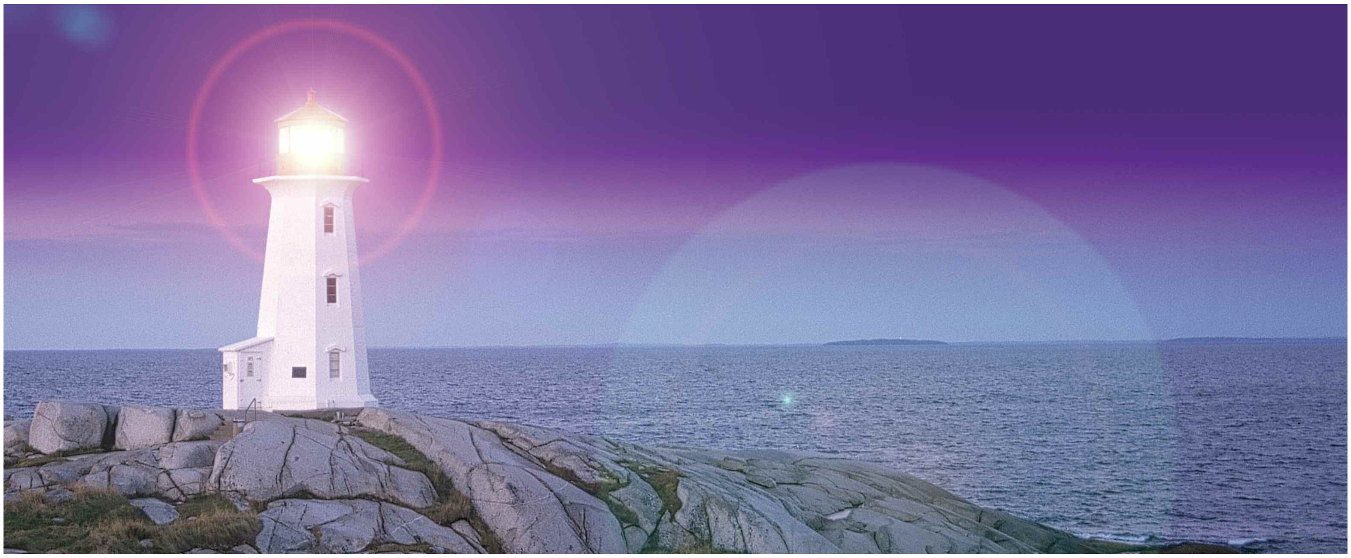

Field Trip A5

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

David Keighley and David E. Brown

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

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NOTES

FACIES HETEROGENEITY IN LACUSTRINE BASINS: THE TRANSTENSIONAL MONCTON BASIN (MISSISSIPPIAN) AND EXTENSIONAL FUNDY BASIN (TRIASSIC-JURASSIC), NEW BRUNSWICK AND NOVA SCOTIA

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SAFETY

For personal and group safety reasons, field trip participants are advised to read and heed the following safety related procedures. The field trip leaders will endeavour to make the trip as safe as possible but can do so only with the co-operation of the participants.

1. Picks and Hammers

Please do not indiscriminately hammer and do not swing the hammer or pick wildly. Use downward blows and make sure no one is standing close to you. Do not pick at rock or till above your head.

2. Suitable Clothing

The weather in the Maritimes can be quite unpredictable at this time of year. Participants should have adequate footwear and clothing that will provide protection against both wet and cold, including hat, gloves, waterproof clothing and boots. Reflective or bright clothing should be worn particularly when visiting roadside outcrops. Traffic vests are useful.

3. Safety Goggles

Participants are advised that safety goggles are recommended. Safety goggles are particularly important when hammering fine grained hard rocks, such as basalts.

4. Hard Hats

Hard hats provide some protection against falling rocks and are strongly recommended. Hard hats can be loaned to participants on request.

5. Falling Rocks

The combination of melting snow, frost action and tides makes spring the most hazardous time of the year along the coastal and road sections in this area. Falling rocks are therefore a major hazard. Every situation should be individually assessed, but the following are useful guidelines:
(a) avoid obviously unstable or overhanging cliffs,
(b) do not hammer above your head or above others,
(c) if a slope must be climbed, do not allow participants to climb while others are below, and
(d) do not undercut unconsolidated cliffs that might slump.

6. Tidal Sections

The extreme tidal range in the Bay of Fundy is responsible for the formation of the excellent exposures we will be visiting during this trip. Conversely, access to these areas is strictly regulated by the tides and a healthy respect for their power is a prerequisite. It is strongly recommended that participants take great care on beach sections adjacent to promontories and headlands noting that they are cut off at high tide. Additionally, many of the intertidal areas are covered in very slippery seaweed, are very bouldery, are very muddy, or any combination of the above. Please take your time and be conscious of your footing. A person will be delegated responsible to bring up the rear of the party and effectively watch for stragglers.

7. Roadside sections

Most of the outcrops visited on day 2 are road cuttings on the main Saint John - Moncton highway. This highway is divided, but traffic can be quite heavy and fast moving. Keep off the paved highway at all times (we should only be crossing to the central divide at one stop). Also keep an eye out for ATVers.

8. Use Common Sense

When in doubt, don't hesitate to ask for guidance.

INTRODUCTION

1) Lacustrine facies & sequence stratigraphy

1.1. Preamble/definitions

One of the main purposes of this field excursion is to demonstrate the variety of sedimentary environments encountered in alluvial-lacustrine sedimentary basins. By way of an introduction, we shall summarize what these environments are using modern day examples, before discussing their recognition and succession in the rock record.

Sedimentary environment:

A sedimentary environment exists where a dynamic suite of physical, chemical, and biological processes collectively operate to produce or modify a body of sediment. "Each sedimentary environment is characterized by a particular suite of physical, chemical, and biologic parameters" (Boggs, 2001, p. 257). Physical processes include aqueous wave and current activity, and gravity processes. Chemical processes include dissolution, hydrolysis and oxidation; biological processes include biochemical precipitation, biological reworking of the sediment, and photosynthesis. The processes acting on the sediment are not themselves preserved in the rock record and must be recognised through observation or sedimentary facies analysis of the sediment or sedimentary rock.

Environmental settings:

A particular suite of physical, chemical, and biological parameters acting on sediment (the sedimentary environment) occurs only at specific locations, or settings, on the earth's surface.

Sedimentary facies:

The physical characteristics (e.g. lithology: grain size, sorting, cement, colour) and sedimentary structures (e.g. cross-bedding style, interface markings) of a body of sediment or sedimentary rock are elements of the response to the aforementioned suite of processes. It is these elements of the response that can be studied to reconstruct the original environmental setting. A lithofacies is a rock unit objectively described, in terms of its lithology and sedimentary structures, for the purpose of providing a subsequent environmental interpretation: for example, a red, rippled sandstone lithofacies. Similarly, contained fossil assemblages can be objectively described as a biofacies, petrophysical properties such as gamma ray response can be objectively described as a petrofacies.

An interpretive facies (Reading, 1978) may be in (1) the 'genetic' sense for the products of a process by which a rock is thought to have formed (e.g. turbidite facies for the products of gravity initiated turbidity currents), (2) the 'environmental setting' sense for the products of the setting where the rock is thought to have formed. An interpretation may draw on any combination of objectively described facies. For the Moncton and Fundy basins we have typically drawn on biofacies, petrofacies, and, primarily, lithofacies.

Individual lithofacies vary in their interpretive value. However, lithofacies encountered vertically and laterally adjacent to one another may be treated as a "lithofacies association" to provide a more robust interpretation (Walther's, 1894, *Law of Facies*, which, is dependent on the nature of the contact between adjacent facies). Interpretations of environmental setting result from (1) comparison of lithology and sedimentary structures in sediments from modern-day settings and similar features (and hence inferred similar setting) in a particular rock unit and (2) an understanding of the environmental parameters that limit the occurrence of particular physical characteristics and sedimentary structures and in what settings these parameters are inferred or known to be found.

Sedimentary basins:

Sedimentary basins are, or were, topographic lows (depressions) within which sediment can, or has, accumulate(d). The occurrence of topographic lows on the scale of 10's of km and larger are usually related to tectonic processes that result in subsidence. Dickinson (1993) considers mechanisms to include crustal thinning, mantle-lithospheric thickening, sedimentary and volcanic loading, tectonic loading, subcrustal loading, asthenospheric flow, and crustal densification. The type(s) of process active, and where the process is active with respect to plate margins, give rise to different kinds of sedimentary basin as classified by Ingersoll and Busby (1995). They include basins at divergent plate margins, where terrestrial rift valleys form due to crustal thinning and sedimentary/volcanic loading, intracratonic basins due to sedimentary/volcanic loading or mantle-lithospheric cooling, and peripheral foreland basins at convergent margins due to tectonic loading. Also, since plates rarely move directly toward or away from one another, there is usually a variable degree of strike-slip movement incorporated in the divergence or convergence. Where transform faults are present, there are often progressively varying zones of transtension (transform-divergence) and transpression (transform-convergence), with basins forming where there is crustal thinning associated with the 'pull-aparts'.

Basin margins and sub-basins

Many sedimentary basins have at least one of their boundaries defined by a faulted margin against which is an uplifted (relatively) basement block (exceptions would typically occur in intracratonic basins).

A sedimentary basin is often viewed simplistically as having a low point or line, towards which sedimentation is directed downslope. While this may be the case for some intracratonic basins, many regions of sedimentation have more than one 'depo-centre', separated from one-another by remnant basement highs or by active areas of intra-basinal uplift. For present day basins accumulating sediment, distinct sub-basins can readily be identified where a sedimentary process is continuous across at least part of the remnant high or area of intra-basinsal uplift. In the rock-record, such a distinction is more difficult.

Depressions separated by uplifts and with fault boundaries undergoing different kinematics are considered separate basins (that regionally may form a complex of basins). Herein, we shall use the term 'sub-basins' for topographic lows that form part of the same structural trend (e.g. have a similar basin-margin bounding fault), and appear to have been accumulating sediment at the same time or progressively in a particular direction.

It is important to remember that, at any particular locality, subsidence mechanisms may vary over time. New fault systems may develop that dissect an existing sedimentary basin, producing smaller independent basins. Other fault-bounded margins may become inactive and buried by continued sedimentation: the structural trend that previously defined separate basins is no longer important and what were separate basins are now part of a much larger whole and at best 'sub-basins' of the larger basin.

Lakes and lake basins

A lake is a body of water that is (almost entirely) enclosed by permanently terrestrial settings. Lake (lacustrine) basins are simply sedimentary basins that contain either a permanent or ephemeral lake in the basin's topographically lowest area. Adjacent to the lake, terrestrial sediments may be accumulating as the result of aeolian, alluvial, chemical, glacial, and/or mass movement processes in fan, floodplain, playa, etc. settings.

Alluvial-lacustrine terminology is discussed on the following pages (see also Figures 1 and 2) using examples from around the world, but our comments are by no means exhaustive in terms of what environmental settings and sediments exist.

1.2. Present day lakes: a general introduction

Lakes vary tremendously in size. The smallest lakes, which are typically rapidly infilled by sediment and therefore short lived, occur in settings such as coastal or alluvial plains or deltas. They are formed by autocyclic processes such as sedimentary damming, local erosion or solution. Larger lakes are primarily tectonic in origin (glacially formed lakes are also widespread today, but likely were less important during much of geological time). Some form in active tectonic areas such as extensional rift valleys (e.g. east African lakes) and strike-slip orogenic belts (e.g. Salton Sea, California), with steep, faulted basement margins and basement 'inliers'. Others form on long-lasting sags in cratonic areas (e.g. Lake Eyre, Australia). Some lakes are deep, others shallow. However, there is no relationship between lake area and depth. Furthermore, neither lake area or depth (nor salinity, organic production, type of sediment being produced) correlates significantly with either tectonic/glacial setting, or climate zone (Bohacs et al., 2000).

Water enters lakes directly from precipitation and from inflowing rivers, groundwater, and hot springs. Some lakes (e.g. the Great Lakes) have a surface outflow by way of a river draining out, across a sill, and away from the lake ("open lakes"). This allows throughflow of water; saline water may be replaced by fresher water. Some 'lakes' (e.g. Lake Maracaibo, Venezuela; Long Reach-Belleisle Bay-Washademoak Lake, New Brunswick) are connected, with periodic inflow, from the ocean and can be considered a transitional feature intermediate with estuaries. Yet other lakes (e.g. Great Salt Lake,-GSL) have no surface outflow and most water loss is through evaporation which concentrates the remaining solution into a salty brine ("closed lakes"). Lake levels in closed lakes may vary on several time-scales (see below), resulting in major shoreline transgressions and regressions. Outflowing rivers provide a buffer from increasing water inputs to an open lake, stabilizing shorelines and water levels.

The composition of lake water is derived primarily from solution of the many different types of sediment and bedrock over and through which surface water and groundwater has flowed, and by local volcanicity and hydrothermal vents (hot springs). Soda-rich lakes are most common in volcanic terrains. The amount of material in solution (salinity) is further governed by the net water budget (input:output), including whether there is fluvial throughflow or marine inflow, and by the ratio of chemical precipitation to solution within the lake.

Temperatures vary between lakes, seasonally within the same lake, and at different depths of a lake. Where a steep temperature gradient (thermocline) develops a colder denser layer may develop at the lake bottom (Figure 1). Where this stratification is stable, lake waters of different depths do not mix. Increased salinities can help maintain this thermal density stratification. Seasonal influxes of fluvial fresh water may flow over saline wedges formed during dry seasons when lake waters undergo greater evaporation and concentration of the brine. Stratified ("meromictic") lakes are most common in the tropics. Lakes of temperate and polar latitudes may have seasonal mixing due to temperature inversions (and so more typically have oxygenated water columns "oligotrophic"). In the winter, the surface waters may cool toward 4^{o}C, becoming denser, with the resultant tendency to sink below warmer waters at depth (Figure 1). Shallow lakes tend to mix regularly ("polymictic").

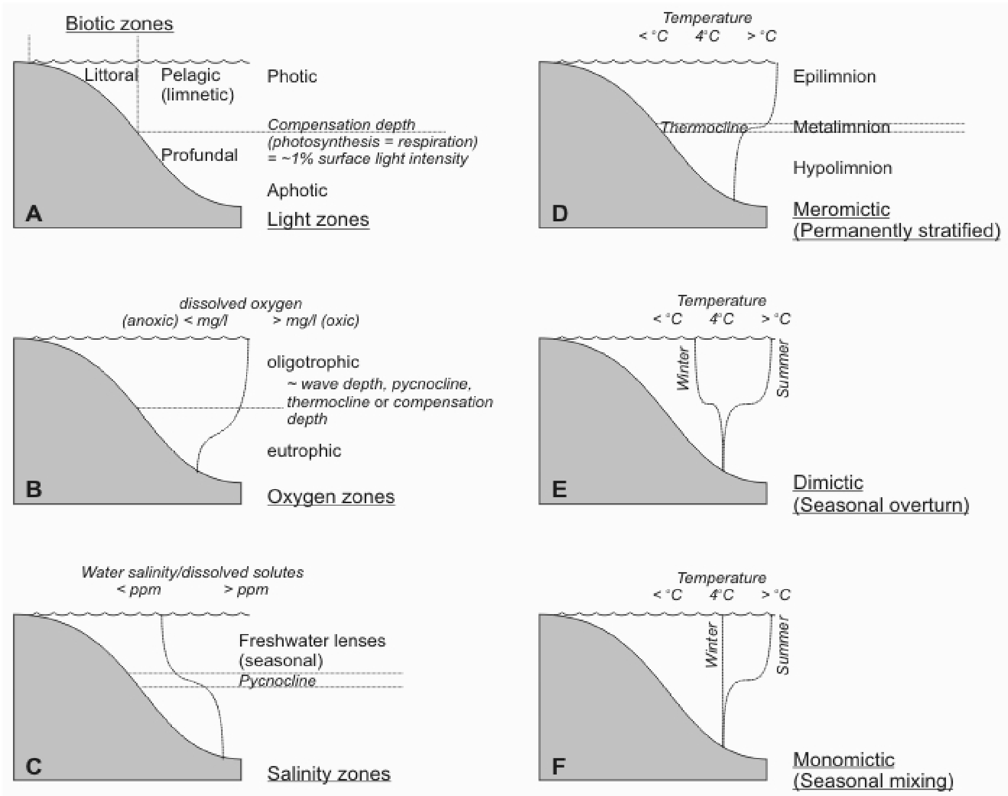


Figure 1. Zonation in lakes (lower zones may not be present in some lakes). **A)** Biotic zonation is related primarily to light penetration. **B)** The depth where there is no free oxygen equates to the compensation depth or base of a zone of water mixing. Stratification of waters, inhibiting mixing, reflects water density. Density can reflect suspended or **(C)** dissolved sediment load, and water temperature. **D)** Warmer water is less dense, forming an upper layer (epilimnion) which typically mixes internally on several time scales. Year-round stratification produces a meromictic lake. Where lake surfaces in the winter cool toward +4C (max. water density) there is a density inversion and cool surface water sinks - lake overturn. If the lake surface further cools and freezes , a further stratification may develop until the surface warms back to 4C in the following spring **(E)**. Such biannual stratification produces a dimictic lake. **F)** If the lake does not freeze there is a tendency for the lake to mix throughout the winter, resulting in only one period of stratification - monomictic lake.

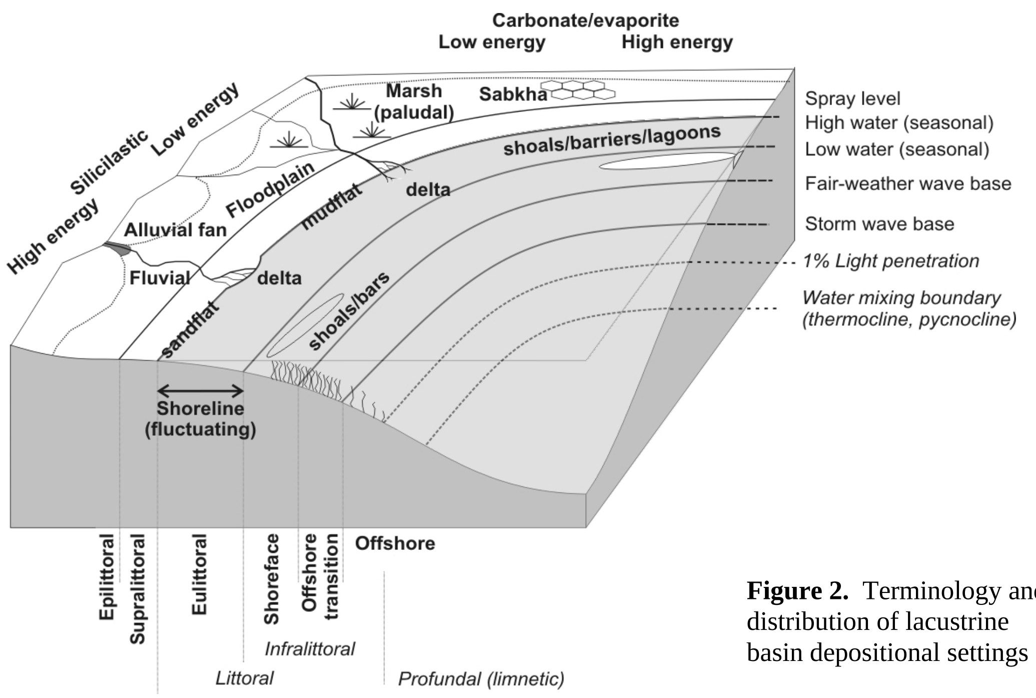


Figure 2. Terminology and distribution of lacustrine basin depositional settings

High nutrient levels and high organic productivity may result in the accumulation of organic debris on the lake floor. Bacterial breakdown of this debris can deplete oxygen in the lower parts of the water column ("eutrophication"), particularly where there is no mixing. The epilimnion (Figure 1) may retain oxygenation, but the lower layer will develop anoxia, retarding organic breakdown by bacteria that require oxygen, and allow for the accumulation of organic-rich sediment.

Lakes have complex patterns of water movement involving currents, waves, and in the very largest lakes, slight oscillations due to tides. Waves aid in the mixing of lake water, re-oxygenation and transporting sediment, and cause larger scale water movements. The importance of waves depends on wave fetch and hence on the orientation of the lake with respect to the prevailing wind. On the shoreface, sediment is moved alongshore and offshore by wave activity.

1.3. Characteristic environmental settings in modern lacustrine basins

Lake basins can, and should, be viewed as potentially containing all of the same environmental settings that could potentially form in a sedimentary basin with marine influence - with the typical exception of intertidal settings (Figure 2). Although no present day lakes are large enough/beneficially oriented to show any appreciable tidal signature (excepting the lakes transitional with estuaries), it should not preclude their possible existence in the past. Accordingly, alluvial fan, braidplain, fluvial floodplain, aeolian sand sea, delta, sabkha, lagoon, beach, shoreface, offshore ramp, slope, or pelagic settings etc., can be encountered.

High resolution seismic profiles, side-scan sonar, and coring of east African rift lakes (e.g. Johnson and Ng'ang'a, 1990; Cohen, 1990; Scholz and Rosendahl, 1990) identify diatom-rich (pelagic) laminated clays in the deep offshore (>100m depth), along with contourite sands, sublacustrine fans (drowned and modified fluvial deltas), and turbidite channels near to clastic sources. Subaqueous talus deposits and fan deltas are also common on the fault-scarp deep-lake margins. Alluvial fans are common on fault scarp ephemeral lake margins. On shallow gradient lake margins (rift flexural/hanging wall block), well developed shorefaces, beaches and deltas are present. These settings are similarly developed in intracratonic, and foreland basin ramp, tectonic settings.

The floors of many closed (and shallow open lakes) have a particularly low and regular gradient (e.g. GSL - averages 0.4 m per km). This is explained as due to wave base or the water surface itself coinciding with, and smoothly grading, the lake bottom during lake low-stands (Currey, 1980). Submerged shorelines can be identified from evidence such as drowned beach ridges and desiccation polygons, the former feature often being reworked to form present day submerged bars and sand waves.

While locally the shore of the GSL laps against outcropping basement that forms irregular, rocky promontories, mostly the shoreline laps against barren gently shelving mudflats of relict lake floor mud which has now been exposed following a drop in lake level. Raised beaches, bars, spits, deltas and terraces, of coarser material not easily eroded by wind and waves, are characteristic features indicating former higher lake levels. Some of the subaerially exposed features have had their finer grained sediment reworked to form eolian dunes, whereas water in low-lying areas and depressions first become isolated from the main lake and then to dry out completely, leaving behind playa-flat deposits (e.g., Bonneville flats, Utah).

As would be expected on any wave influenced beach-shoreface, longshore drift is active in many lakes. Typical features are present in the Great Lakes, such as spits, barriers, and tombolos. Salt marsh may develop extensively in lagoons behind large barriers. The GSL also shows variation in its deltas. Arcuate delta mouth bars have developed at Farmington where wave activity is most pronounced. River activity is most pronounced at the Bear River delta where elongate (birdfoot) levees extend out into the lake.

1.4. Sediment composition in modern lakes

Clastic particles in lacustrine basins are highly variable. Size fractions range from clay to boulder, usually dependent on transport distance from source. Textural and mineralogical maturity is also related to transport distance. The geology (and climate leading to weathering) of a source area also profoundly influences the mineralogy of the sediments, both the detrital grain and clay phases. For example, detrital clays from the GSL comprise 51% K-mica, 39% smectite (montmorillonite) and illite-smectite, and 10% kaolinite (Hedberg and Perry, 1971). Lake Baringo in the rift valley has 60-80% smectite with kaolinite and illite locally also important, relating to provenance (Renaut et al. 2000).

The composition of the dissolved load in lakes mostly reflects the dissolved load of inflowing rivers. For some lakes (e.g. volcanic and rift basin lakes) hydrothermal sources may also be important. Where significant dissolved calcium and carbonate is present, an extensive calcite-shelled invertebrate fauna (typically shallow water benthic - also nektonic, e.g. brine shrimp of the GSL) may populate a lake, with their remains forming beach and bar coquinas. Fecal pellets of the nekton, made up mostly of carbonate and clay, may also cover much of the lake floor. In many other lakes, it is the planktonic, siliceous diatoms that provide a dense pelagic rain to the lake floor.

Carbonate ooidal sands are often found corresponding to the parts of lakes that experience most wave energy (e.g. the downwind southern part of the GSL). The distribution of algal bioherms in the lake parallels that of ooids.

As previously mentioned, the composition of dissolved salts in lakes varies widely, and thus a wide variety of precipitates and authigenic minerals are encountered, particularly in closed lakes. In the GSL, the north arm of the lake contains a dense brine from which precipitates halite ($NaCl$) and, in winter, mirabilite ($Na_2SO_4.10H_2O$). Gypsum is also present, in association with sodium-rich muds, on the lake bottom.

1.5. Lithofacies associations in ancient lacustrine basins

On the basis of comparisons of sediment composition, textures, and structures between ancient and modern, all the aforementioned lacustrine environmental settings have been interpreted as present to a greater or lesser extent in the stratigraphic record. In addition, three recurring lithofacies associations (essentially end-members of a continuum) have long been modelled on the metre to hundreds of metres scale (e.g. Bradley, 1931; Olsen, 1990; Carroll and Bohacs, 1999 - Table 1).

Olsen (1990), investigating the eastern North America Newark Supergroup (including strata of the Fundy Basin) explained the occurrence of these lithofacies associations as reflecting the interplay of basin morphology, climate, and time. Collectively, these factors influenced the magnitude and frequency of lake-level change. Predominantly open lakes over time result in fluvial-lacustrine facies associations (his "Richmond-type"), predominantly closed lakes over time produce evaporative facies associations ("Fundy-type"), and lakes that fluctuated between open and closed produce profundal facies associations (his "Newark type"). Carroll and Bohacs's (1999) equivalent terms are 'underfilled', 'overfilled', and 'balanced fill' successions.

Facies association	Sedimentary structures	Lithologies	Organic matter
Evaporative (Underfilled)	-climbing asymmetric ripples -flat bedded -displacive fabrics -cumulate textures -algal growths (incl. stromatolites) -grainfall-grainflow cross-strata	-siliciclastic mudstone, siltstone, sandstone -precipitates: sulfates, chlorides, potash, bicarbonates -carbonate grainstone, boundstone, flat-pebble conglomerate -kerogenite	-low diversity halophytic biota -algal bacterial organic matter -low to high TOC -hypersaline biomarkers
Fluctuating profundal (Balanced fill)	-flat bedded -symmetric and asymmetric ripples -trough and planar-tabular cross-strata -bio- and pedo-turbation structures -mudcracks -algal growths (incl. oncolites, stromatolites)	-siliciclastic mudstone, siltstone, sandstone -marl -carbonate grainstone, wackestone, micrite, boundstone -kerogenite	-salinity tolerant biota -aquatic algal organic matter -minor terrestrial plant matter -moderate to high TOC -algal biomarkers
Fluvial-lacustrine (Overfilled)	-flat bedded -symmetric and asymmetric ripples -trough and planar-tabular cross-strata -bio- and pedo-turbation structures	-siliciclastic sandstone, siltstone, mudstone -marl -carbonate grainstone, coquina -coal, coaly shale	-freshwater biota -land plant, charophytic and aquatic algal organic matter -low to moderate TOC -terrigenous and algal biomarkers

Table 1: Representative attributes of three major lithofacies associations attributed to lacustrine deposits (after Bohacs et al., 2000)

1.6. Sequence stratigraphy in lacustrine basins

Analysis of the vertical and lateral trends in the sedimentary facies resulting from fluctuating lake level is the discipline of sequence stratigraphy. Sequence stratigraphy, when approached from a sedimentological viewpoint, should be a deductive process. Initially, there is the observation of lithologic, biologic, petrophysical (etc.) characteristics in the rock, and of the contacts between different rock units (e.g. gradational, sharp, erosive). From these observations, facies interpretations can be made. Vertical and lateral trends in the facies successions (e.g. basinward shifts in facies with time) can then be further interpreted on the basis of whether they represent deposition during periods of base level rise or fall. Switches in these trends can be delimited by key surfaces. It is these key surfaces, and the facies trends contained between them, which provide the means of sequence stratigraphic correlation across a basin. Sequence stratigraphy is therefore a "secondary" interpretation. It is only as good as the facies interpretation that preceded it.

Following are definitions of some of the plethora of terms that are in use.

Base level and basin sills

Base level is: "An equilibrium surface... above which a particle can not come to [permanent] rest and below which deposition and burial are possible." (Sloss, 1962). Environmental settings existing above base level are prone to erosion and transient deposition. Settings below base level tend toward net deposition. Base level in nonmarine basins is defined by lake level and the watertable. Direct precipitation over the lake, plus groundwater and surface runoff (sheetflood, channelized fluvial or glacial flow) provides the input (Figure 3A). Output takes the form of evaporation from the lake, biogenic uptake (trees, animals) and throughflow. Throughflow comprises groundwater seeping out of the basin and in some cases, by fluvial outflow where there is drainage spillover at the basin's sill (Carroll and Bohacs, 1999).

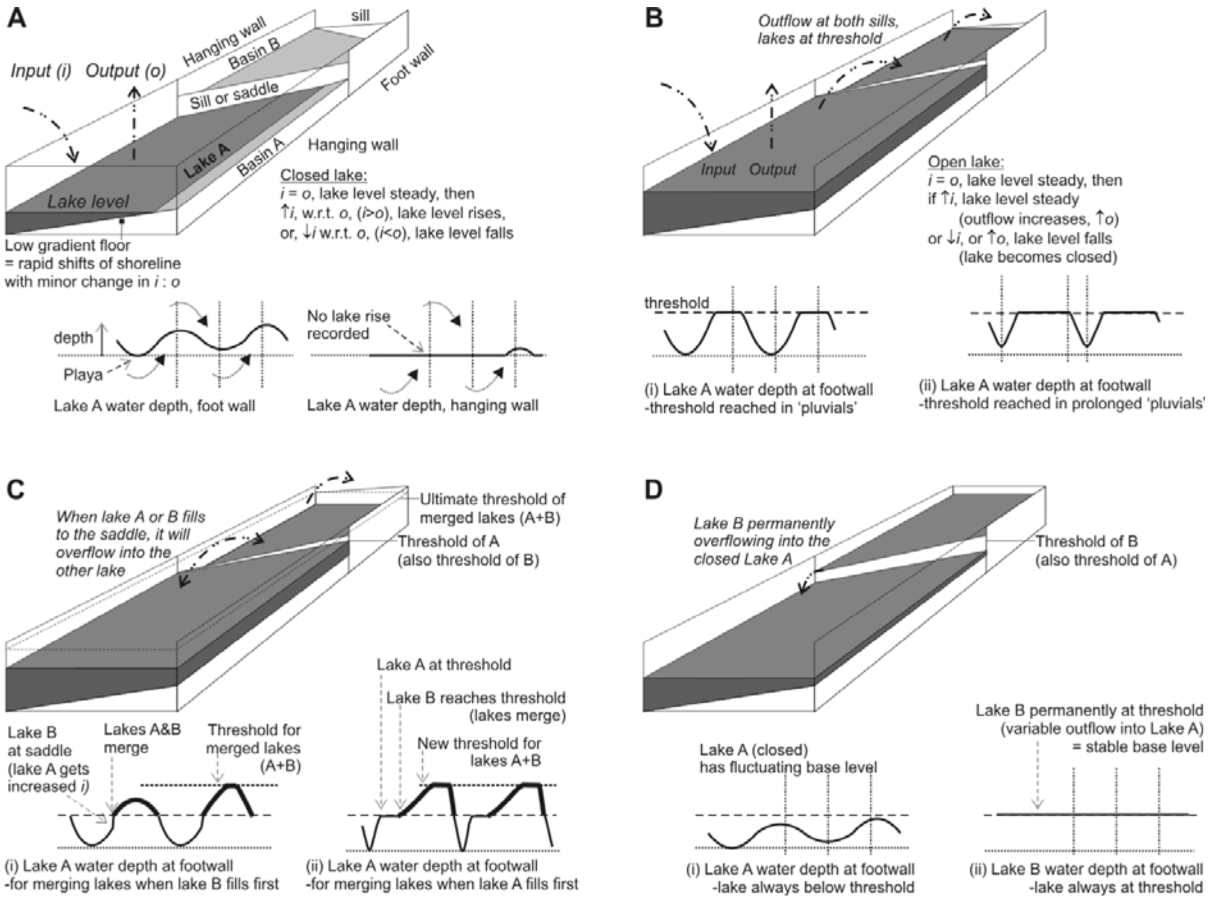


Figure 3. Some theoretical examples of base level behavior in lacustrine basins as a result of selected isolated factors (many others can be envisioned). **A**) Variations in water budget inputs (*i*) and outputs (*o*), with no tectonism involved, results in uniform deepening/shallowing across the basin. The basin and the lake are termed 'closed' unless the lake is at threshold (i.e., has an outflow). **B**) Situation where *i*:*o* increases to bring the lake A to threshold (open lake), resulting in outflow to lake B in an adjacent basin. Lake B is also at threshold, with outflow across its sill, which is at a lower elevation than the sill for lake A. Base level is steady while lake A is at threshold. **C**) The effect of nested basins: lake A that had been at threshold can still deepen when it is able to merge with lake B, because the sill between the two basins is at a lower elevation than any other potential sills. For example, as in case (ii), nested sub-basins exist within the Bonneville Basin of Utah, one of which contains Utah Lake (lake A), currently at threshold. If the adjacent Great Salt Lake (lake B) were to expand, it could merge with Utah Lake forming a new Lake Bonneville (cf. Currey et al. 1984). Note that in case (i), lake B would have a period at threshold, while lake A received greater *i* to bring its lake level up to the elevation of the saddle, after which the level of the merged lakes would rise more slowly than for the single lake. **D)** Adjacent lakes, in similar climatic and tectonic settings may behave very differently if one is closed, the other open. Modified from Keighley et al., 2003.

If a lake lacks an outflow (i.e. it is closed), even slight changes in annual precipitation (input) and evaporation (output) can change base level (Figure 3A). Prolonged time periods with a positive water budget can result in significant increases in the elevation of the surface of the closed lake relative to the deepest point in the basin.

In contrast, if a lake expands to its threshold and there is drainage spillover, further significant base level rise is impossible. Regardless of further increases in input, the lake is balanced by increased spillover output and base level is relatively stable (Figure 3B). Outflow is confined by the geometry of the fluvial channel: it can act as a bottleneck allowing lake levels to increase slightly above the elevation of the sill.

The elevation of the sill above the basin's lowest point will determine how deep a potential lake might become. The sill is always at or above base level and so subject to either erosion by the overspilling lake or to subaerial weathering and mass movement. The geology of the sill primarily determines the rate of erosion at the sill and, in the absence of tectonic rejuvenation, for how long a deep basin with an open lake could persist. The sill of Lake Malawi, east Africa is igneous and the erosion rate is ~8m in 10,000 years; softer rocks forming a sill have greater erosion rates.

A major exception to the sill control on base level exists where the lake in an adjacent basin can fill to the same sill and the lakes merge, the adjacent basin lacking a sill at a lower elevation (Figure 3B, 3C). When the lakes have merged, the lakes may continue to deepen until the next sill elevation is reached (Keighley et al., 2003). In the case where one lake outflows into a closed lake (e.g. Utah Lake into GSL), the open lake may maintain a steady base level over many centuaries, whereas the closed lake undergoes many lake level fluctuations on various time scales (correlation via sequence stratigraphy can become a nightmare in such cases - Figure 3D).

Accommodation:

"The space made available for potential sediment accumulation... [where] in order for sediments to be preserved, there must be space available below base level." Jervey (1988). This 3-D space need not be utilized if sediment supply is insufficient (Figure 4).

Sequence:

Using Lake Bonneville Gilbert Deltas as a case study, Milligan and Chan (1998) suggested that lacustrine sequence boundaries should be based on the established lake level hydrograph rather than the physical stratal surfaces: the succession of strata deposited between successive lowstands of the lake would represent one sequence. This may be a useful definition for the study of modern lakes, where such hydrographic data can be physically measured, but it is not appropriate for analysis of strata from ancient lake basins. In this guidebook, we use adaptations of the geological definitions for a sequence that were first used for marine successions:

"A stratigraphic unit composed of a relatively conformable succession of genetically related strata bounded at its top and base by unconformities or their correlative conformities." (Mitchum et al., 1977).

In this definition, an unconformity is itself defined as:

"A surface separating younger from older strata along which there is evidence of subaerial-erosional truncation and, in some areas, submarine [or sublacustrine] erosion, or subaerial exposure, with a significant hiatus indicated." (Van Wagoner et al., 1990).
A significant hiatus is taken to mean period of time dependent on the scale of the particular study. At a broad level of study, composite sequences are allowed to contain unconformities that, being contained within the sequence-bounding unconformities, are therefore not significant at that scale of study (Emery and Myers, 1996).

The "correlative conformities" in Mitchum et al.'s definition has been a prime source of confusion. Embry (e.g. 1995) considers that the easiest surface to physically identify, and one which comes close to being a time marker, would be a maximum regressive surface (defined below).

Sequence hierarchies

We do not consider the commonly used hierarchy of sequences reflecting global sea level curves (e.g. 2nd order, epoch scale cycles, 1st order era scale cycles of Haq et al., 1988) equivalent or appropriate for lacustrine basins, because base level fluctuates much more rapidly and basin life-span is much shorter; typically a few million years maximum. Similarly, "high-" and "low-frequency" sequences imply a time constraint. Discussion thus refers to the level of detail (resolution) relative to the entire basin fill. However, when considering composite, or nested sequences it must be appreciated that, as in the marine setting, the position of high frequency sequences within the lower frequency accommodation regime can augment (or lessen) the base level rise or fall component of the high resolution sequence.

Sequence boundary:

A sequence boundary is the unconformity and correlative conformity that bounds the bottom or top of the sequence. Proximally, it represents a depositional hiatus, typically of greater temporal duration with increasing distance from the basin centre (Figure 4). Time lines do not cross such a surface. Recognition criteria of sequence boundaries in outcrop section (Van Wagoner et al., 1990) include regional:
- subaerial-erosional truncation
- laterally correlative subaerial exposure surface marked by soil or root horizons
- basinward shift in facies across the boundary (facies dislocation)

Flooding Surface:

According to Van Wagoner et al. (1990), this term reflects a "surface separating younger from older strata across which there is evidence of an abrupt increase in water depth [upwards]." Though defined by these authors for marine flooding surfaces, the term is equally applicable to lake transgressions. In subaqueous settings, the surface approximates to a time line. In terrestrial settings, beyond the limits of the transgression, the surface is difficult to recognise (possibly reflected in soil type, fluvial morphology, etc.).

In this guide, DK also uses the term for the surface corresponding to the base of the transgressing lacustrine strata, and thus for a diachronous surface. This is because subaerially deposited strata dominate in the investigated section and so the diachronous surface is much easier to identify than a time line in the terrestrial succession. Elsewhere, such a diachronous type of surface has been termed a ravinement or a transgressive surface of erosion. To DK's knowledge, such features have not been categorically recognised in lacustrine strata, but recognition criteria would include:
- transgressive lag of shell fragments, rip-up clasts, or pebbles on the transgressed surface
- landward shifts in facies, such as contact between laterally extensive strata of lacustrine origin over strata of terrestrial origin.
- within lacustrine strata, at a sharp contact of shallow water facies (sandstones, oolitic carbonates) on deeper water facies (mudstones, organics, precipitative non-coating, carbonates).

Maximum Flooding Surface:

Maximum Flooding Surfaces are identified by a major facies dislocation across a surface upward into strata interpreted to represent the deepest water strata present within a sequence. At this time, the source of clastic input has been moved furthest toward the hinterland, potentially limiting clastic deposition to the basin margin. Deep water carbonates and oil shales have the potential to develop. Overlying strata should represent a facies succession reflecting regression (Figure 4).

Note again that recognition of the deep water strata is dependent on accurate facies analysis. Mudstones, organics, micrites may accumulate not only in offshore deep waters but also in shallow water, protected settings such as lagoons, interdistributary bays etc. Such deposits may be entirely autocyclic and not be related to a high base level. Useful recognition criteria for the deep water strata include:
- lateral extent.
- lack of terrigenous macro- and microfossils (palynomorphs) or type III kerogens.

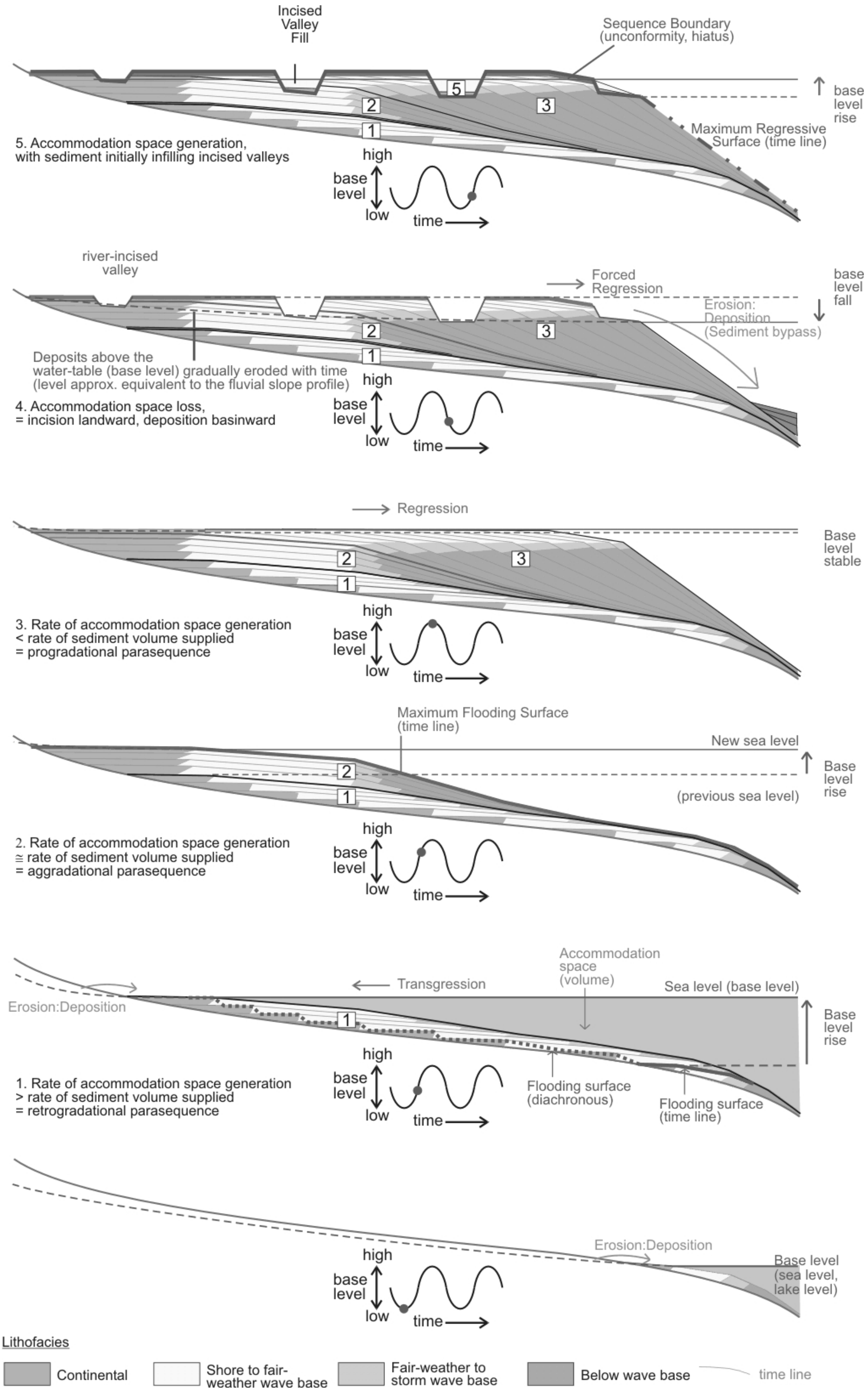


Figure 4. The basic sequence stratigraphic model (evolves from bottom to top)

Maximum Regressive Surface:

Toward the basin centre, deposition may remain below base level throughout a transgressive-regressive cycle, and no sequence boundary may develop. However, base level fall would be represented by a trend in the facies from deep water to shallower water. When base level rise commences, this trend should be reversed. The contact where the facies trend reverses will correspond to the onset of transgression and, shoreward, it will pass into the (still developing) sequence boundary (Figure 4). It may, but need not, correspond to a flooding surface.

Parasequence:

"A relatively conformable succession of genetically related beds or bedsets bounded by marine- [or lacustrine-] flooding surfaces or their correlative surfaces." Van Wagoner et al. (1990).

Parasequence Set:

"A succession of genetically related parasequences forming a distinctive stacking pattern bounded by major marine- [or lacustrine-] flooding surfaces and their correlative surfaces." Van Wagoner et al. (1990). See Figure 4.

Systems Tracts:

Contemporaneous, three-dimensional lithofacies assemblages that subdivide a sequence on the basis of their position within the sequence and the type and distribution of the contained parasequence set(s) and bounding surfaces have been identified as systems tracts in marine basins (cf. Van Wagoner et al., 1990). In the study area, some tracts similar to those encountered in marine settings have been observed. They include:

- Transgressive Systems Tract (TST): between the diachronous, transgressive, flooding surface and the maximum flooding surface.
- Highstand Systems Tract (HST): between the maximum flooding surface and the overlying sequence boundary (this tract includes strata deposited during regression and, potentially, forced regression).
- Lowstand Systems Tract (LST): between the sequence boundary and overlying flooding surface (the tract will include terrestrial sediments deposited at the time when, basinward, the lake had already commenced transgression).

1.7. Low-resolution lacustrine sequence-stratigraphic models

Models for lacustrine sequences are still in their infancy and liable to be significantly modified in future years. They should be considered flexible to new data and ideas.

Tectonic Models

Low resolution tectonic models readily illustrate the conceptual differences that exist in comparison to basins influenced by global sea level (e.g. Lambiase, 1990; Schlische and Olsen, 1990). Basins isolated from marine influence may form in various orogenic regimes: pull-aparts in transtensional/transpressional settings (e.g. Moncton Basin), rift valleys in extensional settings (e.g. Newark - Fundy rift complex); block faulted basins in compressional regimes (e.g. Laramide basins), or post-orogenic sags.

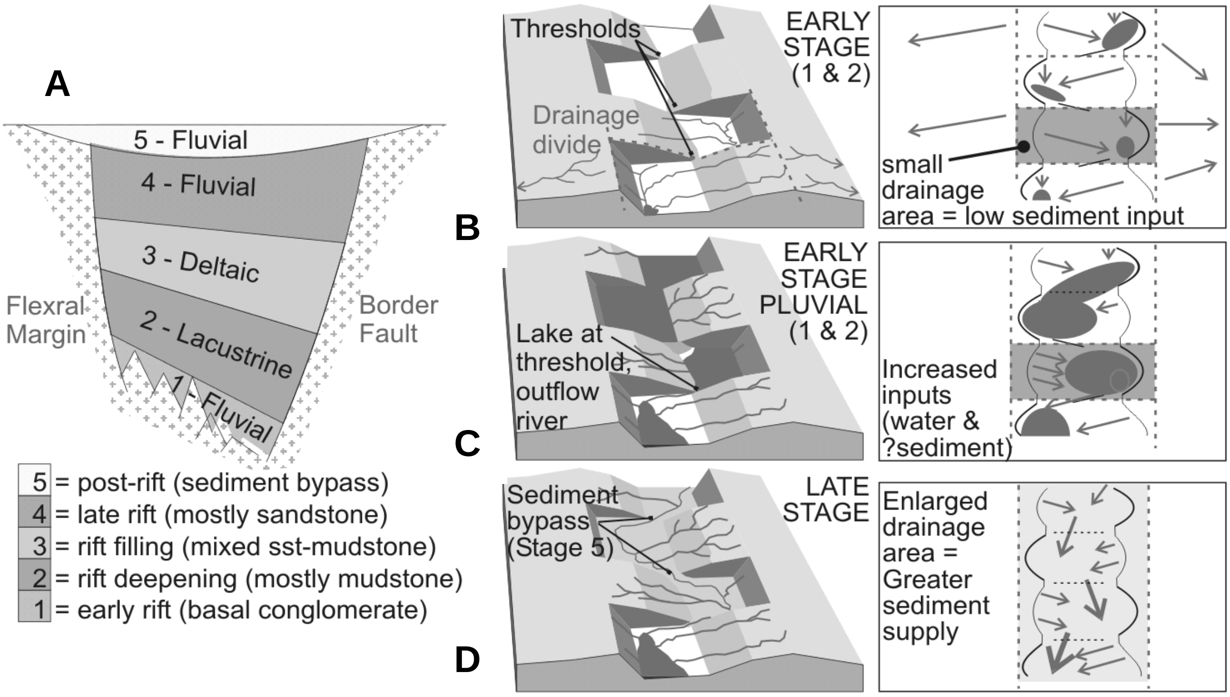


Figure 5 Low resolution tectonic sequences. **A)** Cartoon cross-section across a rift basin, showing Lambiase's (1990) 5 stages of basin fill. **B)** Model for 'stage 2' sediment-starved basins and 'closed' lakes. **C)** Model for the same 'stage 2' of basin fill, but during pluvials, - note that there are some 'open' lakes. **D)** Model showing successive lake basins in different stages of fill: 'stage 5' bypass in the background, 'stages 3 and 4', in the foreground.

None of these basins produce the classic continental shelf/shelf margin profiles upon which sequence-stratigraphic principles were originally developed. In rift and block faulted basins, basement margins will be steep and may exceed clinoform angle. Basement floors are typically low angle, ramp-like, and dipping toward the footwall block. A tectonically active footwall margin can be considered a growth-fault margin. Bedrock shelfal margins are absent.

Lambiase (1990) proposed a five-stage, low resolution model for the filling of nonmarine rift basins (Figure 5), that we consider similarly applicable to any long-lived (>M.yr), actively subsiding, intermontane basin. For any tectonically active nonmarine basin to develop, subsidence of the basin floor, or basin margin uplift, must initially produce volumes of accommodation at a higher rate than supplied sediment can infill this volume.

During this 'growth phase' of the basin, predominantly coarse-grained alluvial deposits (stage 1) initially infill major topographic lows until the basin develops internal drainage. At this point (stage 2), alluvial deposits may continue to form around the basin margin, but fine grained lacustrine deposits tend to dominate in the basin center. Sediment starvation in these earlier stages of the basin-fill succession is favoured because drainage areas are small and fragmented, with limited axial sediment supply. Eventually, the growth phase will end. This may be due to a reduction in the tectonism that periodically deepened the basin, or because of increasing sediment supply. In the latter case, the catchment area becomes significantly enlarged due to infill and subsequent bypass of adjacent, higher elevated basins. The result in both cases is net loss of accommodation. Where there is a deep basin present, thick delta front and delta top sediment packages may develop (stage 3) as the basin infills. As these deltaics extend across the infilling basin, typically in an axial direction, floodplain and fluvial deposition expands landward of the deltas (stage 4). On a temporal scale, basin fill may be relatively rapid compared to the period of basin growth. When the basin has filled, sediment then bypasses (stage 5).

Given identical hydrological input and output throughout the entire history of the basin (hypothetical!), the later stages of basin fill (where sediment supply rate exceeds accommodation space generation) should be characterized by more extensive progradational architecture and its subsequent preservation. When the basin is inferred to be sediment starved, retrogradational and aggradational architecture is favoured (stage 2, accommodation space generation exceeds sediment supply rate).

Lambiase (1990) and Olsen (1990) both point out that basin rejuvenation due to renewed tectonic activity can result in a repetition of stages to produce a dual cycle (or potentially triple cycle etc.), such as in the Newark Basin.

Climate Models

The duration of any hydrologically open or closed lake will exert a primary control on the architecture of the basin fill (Olsen, 1990; Scholz et al., 1998; Carroll and Bohacs, 1999; Bohacs et al., 2000; Figure 6). As previously illustrated (Figure 3C), open lakes promote stable base levels. Closed lakes are prone to fluctuating base levels and thus fluctuating accommodation.

If the basin exists under a primarily wet climate ('overfilled' lake basin of Carroll and Bohacs, 1999), clastic sediment accumulates around the basin margins forming terraced aprons and highstand deltas, which may have high relief if the sill permits deep lakes to form. With a periodically drier climate, base level can drop well below threshold and fluctuate rapidly ('balanced-fill' lake basin of Carroll and Bohacs, 1999). Shorelines and associated facies similarly fluctuate in their position and, depending on the duration of the dry period, partial to complete incision and reworking of the apron deposits may occur (Olsen, 1990). An 'underfilled' lake basin occurs where lake level rarely reaches threshold (Carroll and Bohacs, 1999).

Periodically dry-wet climate cycles have been inferred as having occurred on many different time scales to produce cyclic lacustrine successions. Olsen (1986) considers sedimentary cycles in the order of 1.5 to 35 metres thickness in the Newark Supergroup to reflect Milankovitch-type climate changes primarily related to the 19,000 and 23,000 year precession (axis rotation) cycle and 41,000 year obliquity (axial tilt) cycle. The lower frequency eccentricity (elliptical-circular) cycles at105 ka and 410 ka cycles are also likely important.

It is important to re-emphasize that a lake exists in one of two states, open or closed. The three basic lithofacies associations that are described reflect the relative importance of the open/closed state of the lake *only* at the scale of tens to hundreds of metres of sediment thickness: predominantly open, or closed, lake conditions during deposition of most of the hundreds of metres of sediment, or a common fluctuation between the two states. At higher (metre-scale) resolution, the associations may not be apparent.

Also, since many lacustrine basins contain sediment several kilometers thick, there is always the potential to see a progression upsection from a predominance of one lithofacies association to another.

Interaction of Climatic and Tectonic Controls

Tectonic and climatic controls influence base level to varying degrees, depending on their interplay. When the lake is at threshold, climate change causing increases in precipitation can cause greater sedimentation rates but little further rise in base level, whereas tectonism around the basin margin can raise or lower the sill and hence base level.

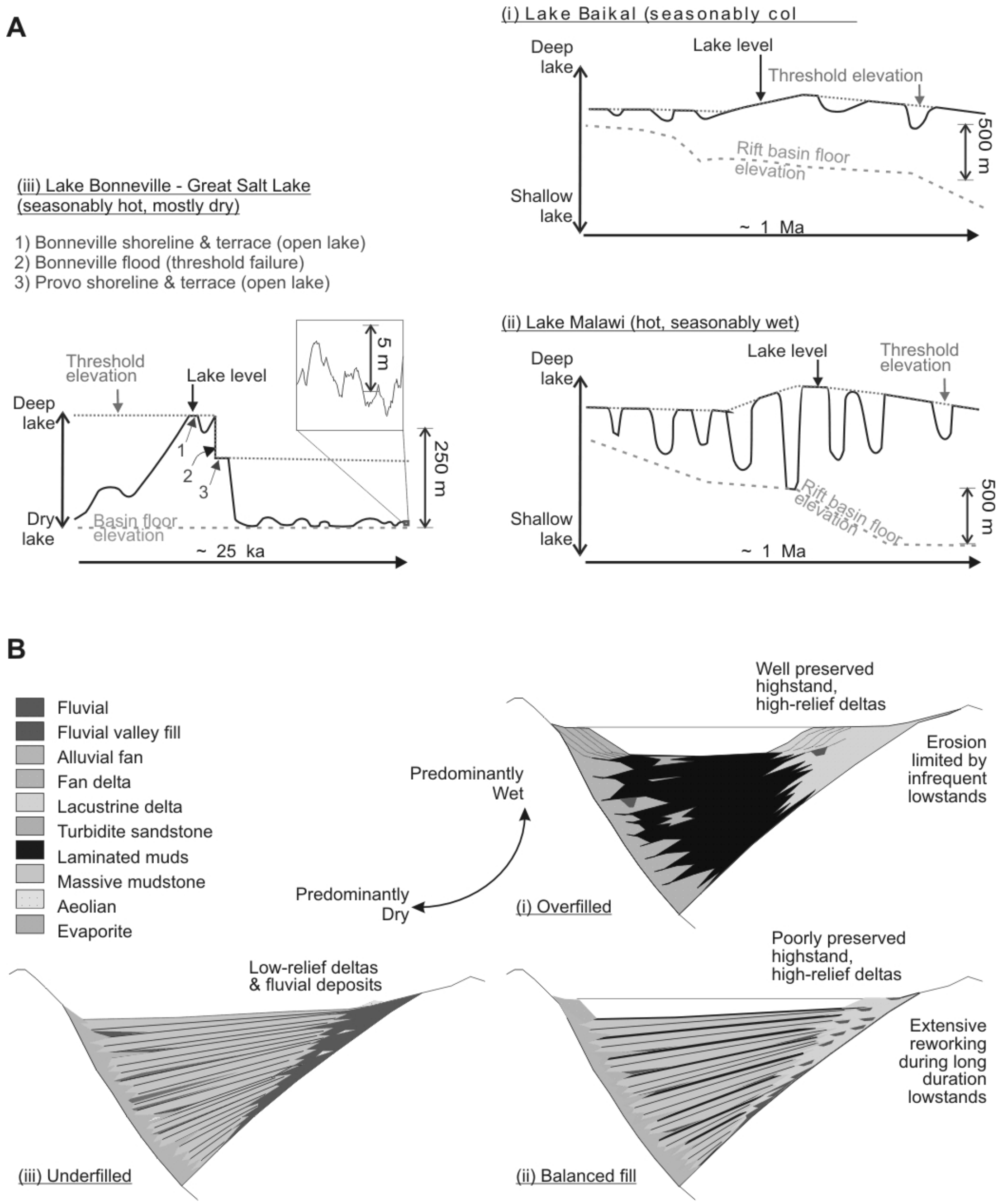


Figure 6 Climate influence on lake -level curves and sequence architecture. **A)** Lake level curves (redrawn from Currey et al., 1984; Scholz et al., 1998) and **(B)** sequence architectures (Olsen, 1990) for early rift phases in i) high-latitude, persistently wet (overfilled) systems, ii) tropical (balanced) systems, and iii) sub-tropical, semi-arid climate (underfilled) systems.

Erosion can lower the sill (Figure 7A). Tectonism can also introduce a new sill and new sediment supply points, while tilting of basin-floor fault blocks can cause a sudden relocation of shorelines and depocenters (Figure 7B, C). For example, if the shoreline (lake level) is on the downthrown side of the hinge-line, the shoreline suddenly relocates toward the downthrow side and a deeper lake forms (Strecker et al., 1999). Such tilting can result in an open lake becoming closed if the volume of the basin (space below threshold elevation: potential accommodation) is increased by such tectonism. A new threshold stage then can be gradually re-established with time *and* a positive water budget.

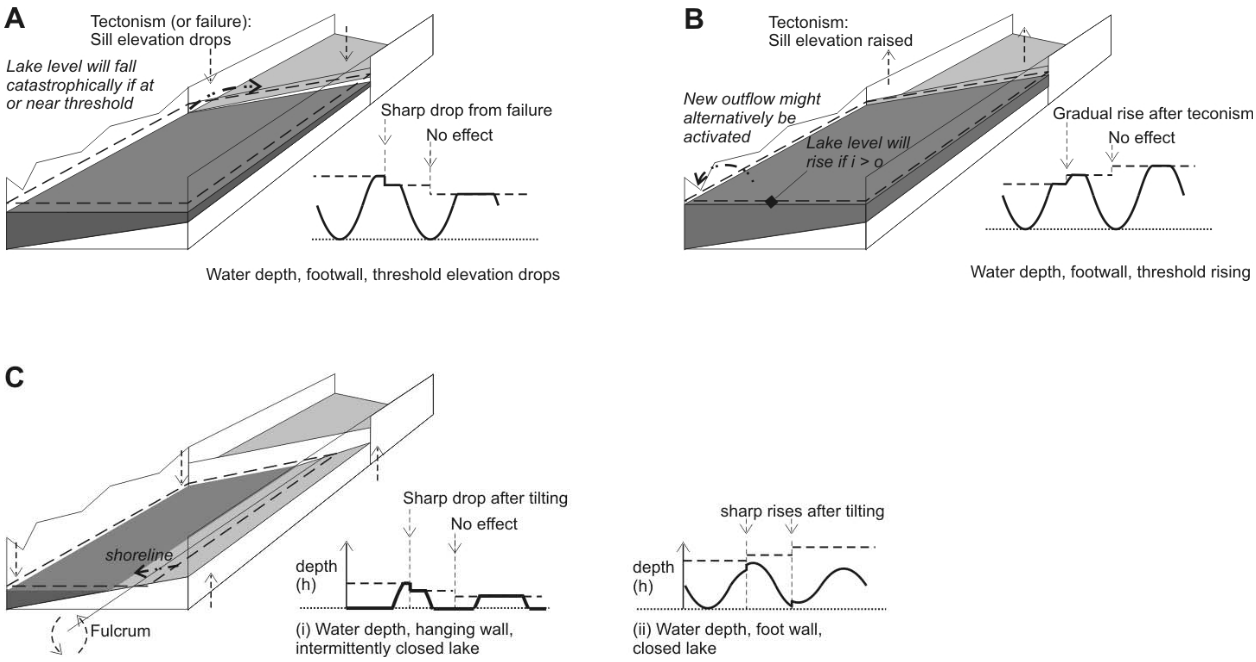


Figure 7. Interplay of tectonic and climate controls influencing base level. **A**) and **B**) Lake levels can also be affected by basin margin tectonism, or by erosion of the sill. Such marginal tectonism or erosion results in uniform deepening/shallowing across the basin if the contained lake is at threshold. Any drop in lake levels will be sudden, since volume available is reduced. For example, the threshold of Utah's Pleistocene Lake Bonneville dropped catastrophically following an erosion-related collapse of its sill (Currey et al. 1984). A rise in the sill will result in a gradual rise in lake level to the new sill elevation since the increased volume has to be filled by subsequent inputs to the lake. Note that tectonic elevation of the active sill may activate a new sill elsewhere which is at a lower elevation. **C**) Internal basin tectonism, whereby the block underlying the lake tilts relative to the marginal blocks (the fulcrum is within the basin), results in the lake shallowing at some locations (situation i), while other locations display a deepening (situation ii). Modified from Keighley et al., 2003.

Climate influenced base level change is more pronounced in balanced and underfilled lake basins, where the water budget is periodically negative. In underfilled lake basins, tectonic influences within the basin are limited to changes of orientation of the basin floor fault block(s) that suddenly shift shorelines and depocenters (Figure 7C). Ongoing basin subsidence or uplift of the sill simply increases the potential accommodation that might be made available following renewed rise in base level. Uplift of basin margin blocks might indirectly raise base levels or increase supply by affecting climate: promoting precipitation over the higher peaks, and hence increasing runoff and sediment supply. Additionally for underfilled lake basins, sediment deposited in a lake displaces water volume, and so the case can arise where reductions in aqueous input can still be reflected in a relative rise in lake level (Einsele and Hinderer, 1997, 1998).

2) Structure, lithostratigraphy, and petroleum of the Moncton Basin (and adjacent basins)

2.1. Structural Setting

Atlantic Canada is underlain by various juxtaposed terranes that form the basement to the Upper Palaeozoic and Mesozoic sedimentary basins to be discussed in this guidebook. The most recently proposed model for the accretion of these terranes was put forward by Barr et al. (2002, 2003) and Percival et al. (2004) (Figure 8). Briefly, the St Croix terrane is considered part of Ganderia (van Staal et al., 1996) while the Mascarene terrane is considered a back arc and the Kingston terrane a volcanic arc, both of which may have developed on the older New River terrane (Barr et al., 2002, 2003). Collectively they form the composite Gander zone. The other peri-gondwanan terranes are the Caledonia terrane and the Brookville terrane, which form part of the classical Avalon zone (Barr et al., 2002). Successively southward, the Gander zone terranes and then the Avalon zone were accreted to Laurentia during the compressional Early Devonian Acadian orogeny. Transpressive motions, related to juxtaposition of the Avalon and Meguma terranes with previously accreted more inboard terranes continued through the Late Devonian and into the Carboniferous (Barr et al., 2002).

The post-Early Devonian stratigraphic succession contains regionally extensive, but relatively thin (typically less than 250 m in New Brunswick) deposits of Pennsylvanian age overlying a thick, but more localized succession of Late Devonian and Mississippian age. This latter succession is mapped along numerous linear belts in southern New Brunswick and northern Nova Scotia (Figure 8). These belts follow the regional southwest-northeast fabric developed in the earlier Palaeozoic rocks through collisional tectonics.

The nature and distribution of these linear Devonian-Carboniferous belts has been best (but by no means completely) explained by Bradley (1982) who related the features of intracontinental transform settings (e.g. McKenzie, 1978) to the region. Following continental collision in the Devonian, the northern Appalachians were influenced by right-lateral strike-slip along the main transforms (Figure 8; Yeo and Gao, 1987; Nance, 1987). Adjacent to the boundary were contemporaneous areas of transtension, forming pull-apart basins, and transpression producing depositional unconformities and deformation. A good analogy is the San Andreas fault system in the vicinity of the Salton Sea, Range Basin, and Los Angeles Basin in southern California (e.g. Wilson, 1962; Crowell, 1974). Transtension was accompanied by increased heat flow, locally with alkaline to tholeiitic volcanism. As transtension diminished, the dropping of isotherms caused periodic regional subsidence and the production of an enlarged sedimentary basin, which buried most of the rapidly sedimented rift basins with a blanket of more mature, more slowly deposited sediment. A good analogy is the North Sea (McKenzie, 1978).

Williams (1974) considered the pull-aparts to be sub-basins of his SE Maritimes Basin which, along with the New Brunswick Shelf comprised his Maritimes Basin (synonymous with the Magdalen Basin of Bell and Howie, 1990; Maritimes Carboniferous Basin of Roliff, 1962; and Fundy Basin of Bell, 1958; Belt, 1968a). However, if the entire Upper Palaeozoic basin fill is to be included within one term, it is suggested the term 'Maritimes Basin Complex' be used (Figure 8). This would emphasize the evolution of the region from numerous smaller scale basins that were likely separate depositional entities at least periodically in late Devonian and Mississippian times (e.g. Cumberland Basin, Magdalen Basin *sensu stricto*, and Moncton Basin), to one large basin depocentre in the Pennsylvanian, the Greater Maritimes Basin.

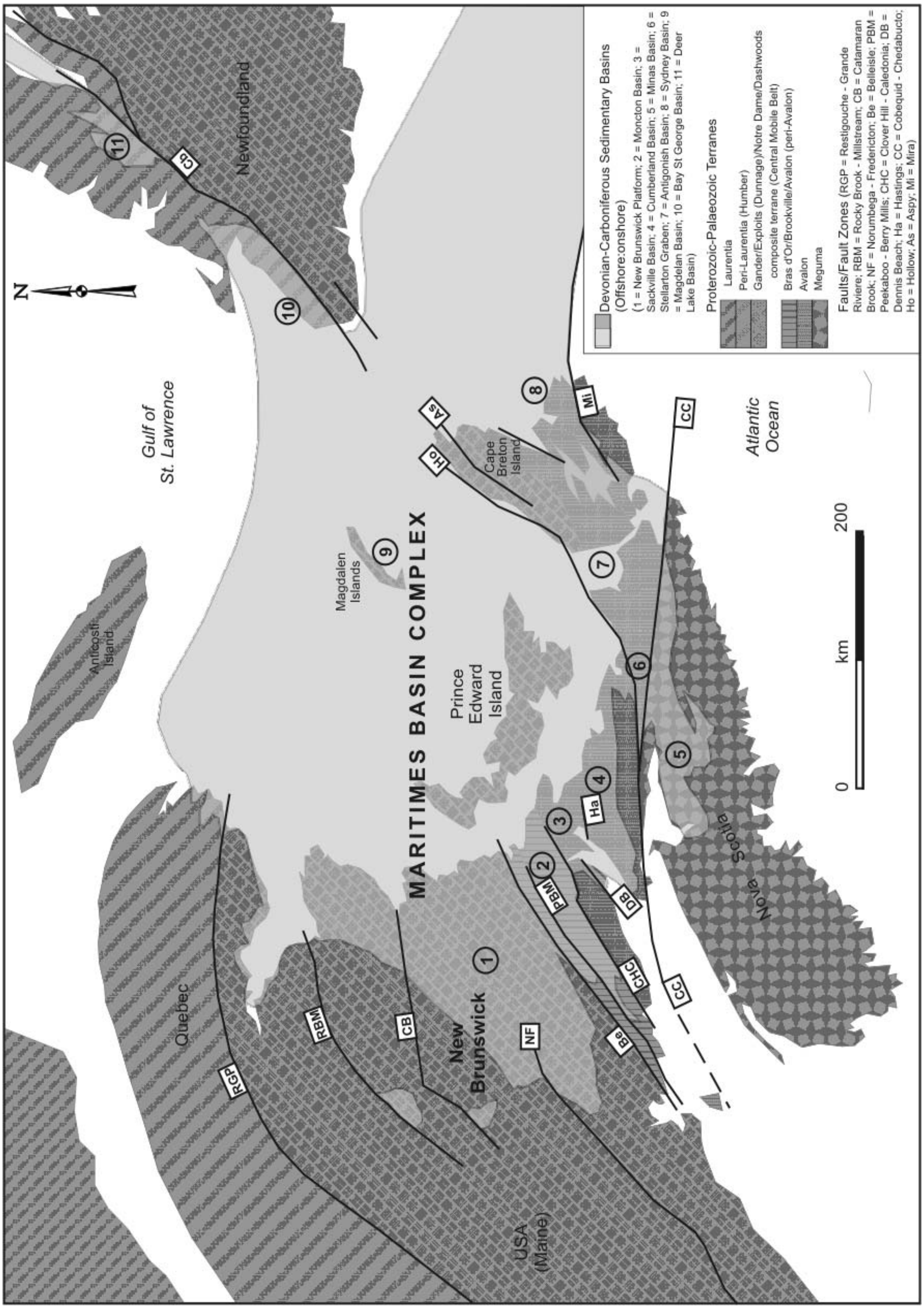


Figure 8. Major structural elements of the Maritimes Basin (adapted from Gibling et al., 1991; Percival et al., 2004)

Late Devonian-Mississippian strata of southern New Brunswick are mostly preserved in the 200km long, northeast trending, wedge-shaped Moncton Basin. The presently defined boundaries of this basin, being marked by faults that also delineate the aforementioned Neoproterozoic-Palaeozoic terranes, are not necessarily the location of the basin-bounding uplifts during sedimentation, because the nature and timing of activity on these faults is poorly constrained and much debated. The main faults are the Belleisle - Lubec and the Peekaboo (Kennebecasis) - Berry Mills fault zones in the north, and the Clover Hill - Caledonia Fault zone in the south (Figures 8 and 9).

The relationship of the Cocagne Graben with the Moncton Basin to its south remains uncertain. Specifically, the nature of the 'basin-bounding' fault remains a challenge, as does the nature of Horton Group strata that include rocks uncritically lumped into the Albert Formation. Do they represent (coeval?) deposits from the same depositional basin as the Moncton Basin or an adjacent sub-basin or basin? On the opposing southern margin of the Moncton Basin, similar questions must be asked regarding its linkage with the Sackville Basin that straddles the New Brunswick - Nova Scotia border.

In the Moncton Basin, ongoing structural studies have resolved a late Tournaisian deformation event with the architecture of a fold-thrust belt in the east of the basin (Park et al., 2005), and two basin inversion events within the Tournaisian are identified in the west (Wilson, 2003; 2005). A later (end Mississippian) event produced reverse/thrust faulting that likely includes allochthonous basement blocks (Keighley and Gemmell, 2005), while the tectonic history is concluded by Pennsylvanian sinistral displacements and reverse faulting, followed by Mesozoic normal faulting (Wilson, 2002). Recognition of the various stages of deformation in the Albert Fm. are further complicated in the field by widespread soft-sediment deformation in the form of slumps (Wilson et al., 2004).

2.2. Lithostratigraphy

The precise stratigraphic relationships between rocks of all ages in southern New Brunswick are still incompletely understood. This continuing uncertainty is a result of several factors, including the following:

1) The region is well vegetated and the presence of Quaternary drift further limits the amount of outcrop available for study.
2) Many of the units have undergone several episodes of deformation, juxtaposing units in faulted contact.
3) There appear to be several stratigraphically distinct red-bed units that cannot be easily distinguished lithologically, particularly since they contain little material of biostratigraphic use.
4) There has been inconsistent and cross-usage of stratigraphic terms, particularly with respect to what is now accepted as a valid lithostratigraphic, biostratigraphic, or chronostratigraphic unit (cf. North American Commission on Stratigraphic Nomenclature, 1983), and imprecisely defined lithostratigraphic units.
5) Biostratigraphic subdivision is complicated by the reworking of Devonian-Carboniferous microfossils within the succession.

Rocks of the early Devonian and older are herein considered part of the 'basement' complexes, and include rocks assigned to the Caledonian, Brookville, Kingston, New River, Mascarene, and St. Croix terranes. It is the rocks of Late Devonian and Carboniferous age that form the sedimentary deposits of the Moncton Basin.

In southern New Brunswick, these sedimentary strata have been summarized most recently by St. Peter (1992, 1993) and St. Peter and Johnson (1997). These authors originally identified four unconformity bounded, tectonostratigraphic successions (or "allocycles"), but the ongoing work now indicates that at least five successions can be correlated regionally (Figure 10).

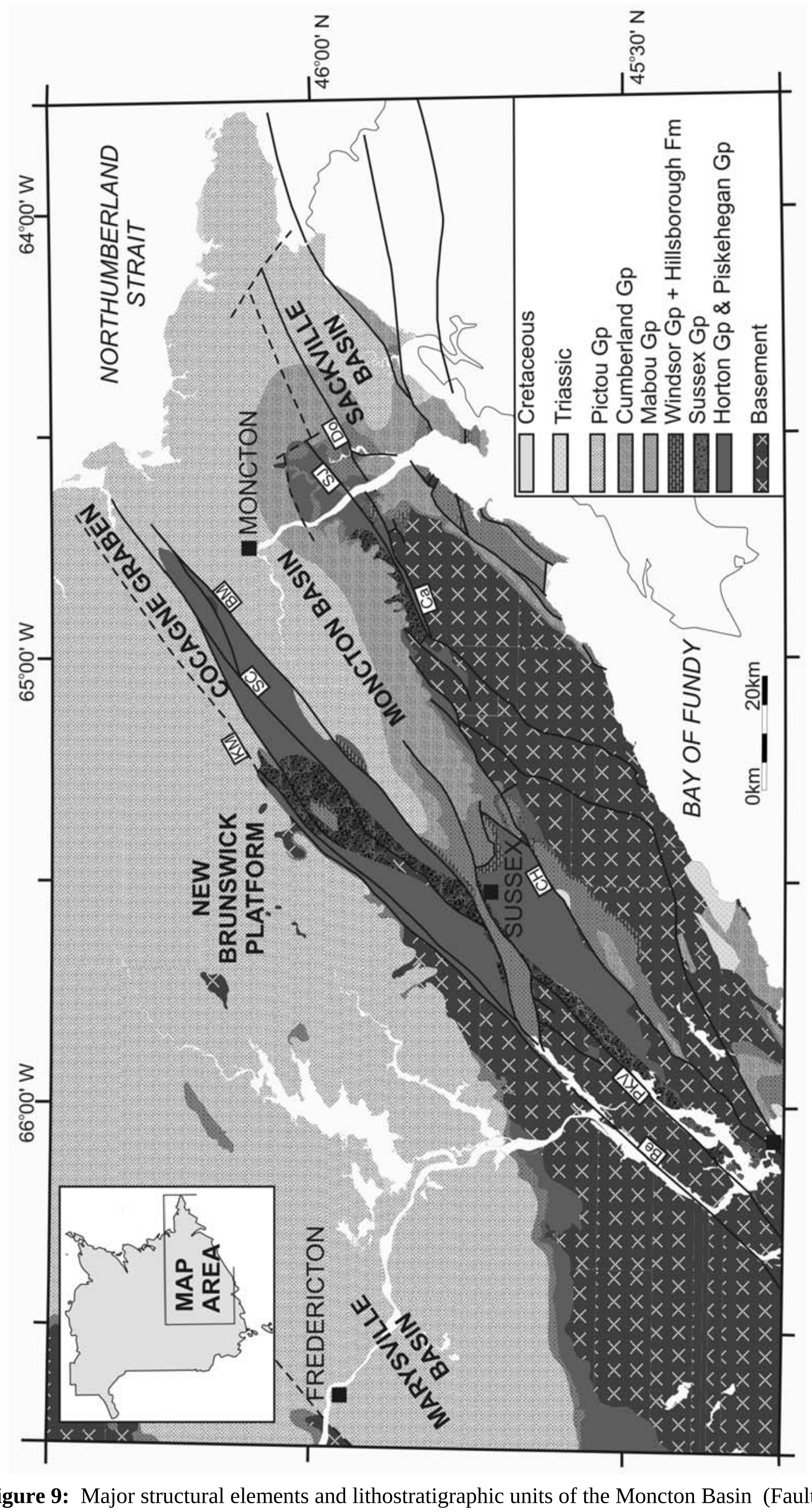


Figure 9: Major structural elements and lithostratigraphic units of the Moncton Basin (Faults: Be = Belleisle, BM = Berry Mills, Ca = Caledonia, CH = Clover Hill, Do = Dorchester, KM = Kierstead Mountain, PKV = Peekaboo or Kennebecasis Valley, SC = Smiths Creek, SJ = Saint Joseph

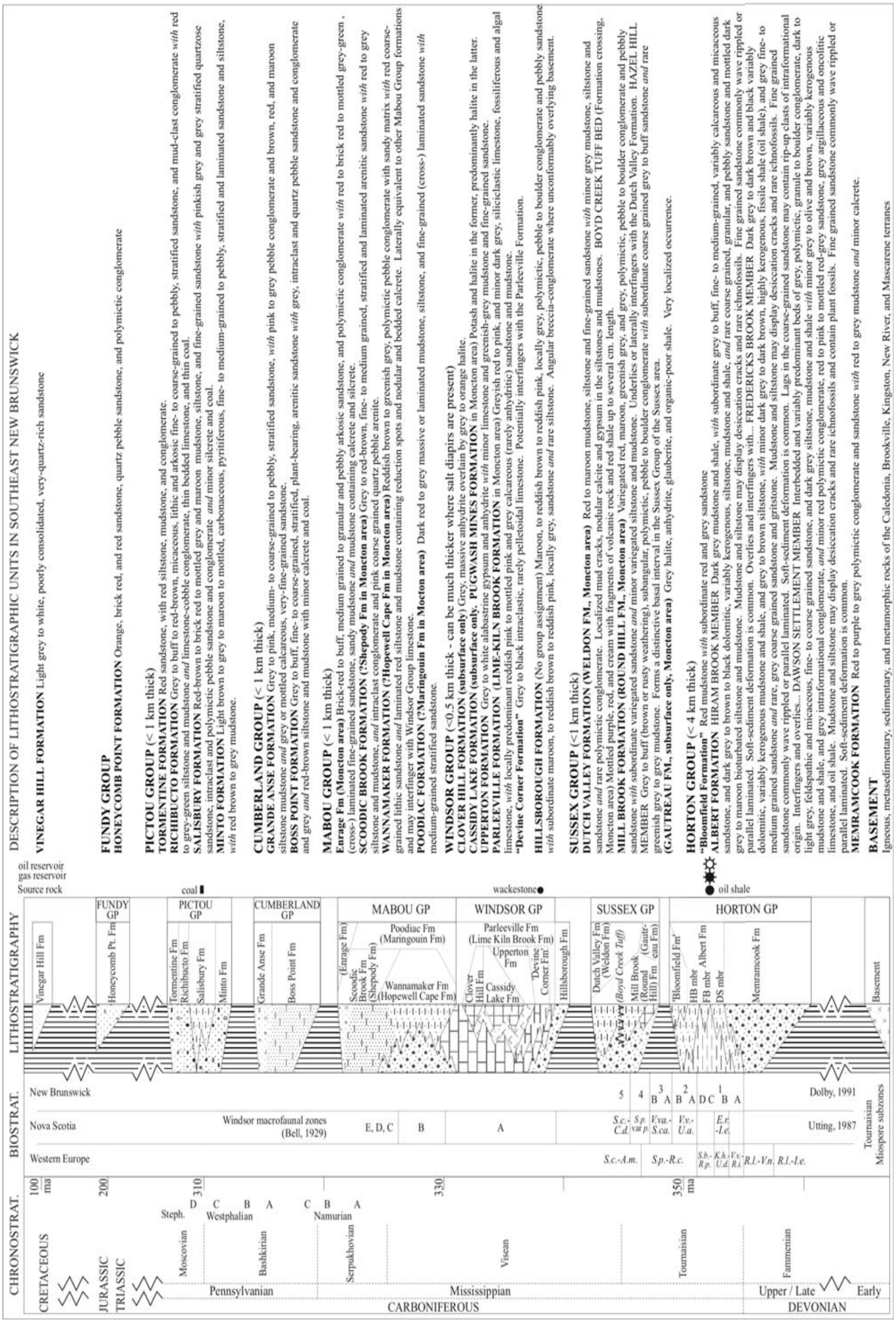


Figure 10: Stratigraphy of the Moncton Basin (includes data from Bell, 1929; Dolby, 1991; Gradstein et al., 2004; Utting, 1987, St. Peter & Johnson, 1997)

The lowermost (Fammenian to Tournasian) succession comprises red-bed strata that interfinger and enclose a unit of mostly grey coloured strata. The basal red-bed unit, assigned to the **Memramcook Formation**, consists mostly of conglomerate, together with gritstone, sandstone, and mudstone that are more common upsection. The grey-bed unit is mostly of shale, but includes oil shale, sandstone, gritstone, conglomerate, and limestone. This latter unit, the **Albert Formation**, is host to the source and reservoir rocks of the McCully gas field, east of Sussex, and the old Stoney Creek oil and gas field south of Moncton. Historically, the red-beds that interfinger and conformably overlie the Albert Formation have been included in the Weldon Formation, although recent work indicates that a new lithostratigraphic name for these strata is required (informally, the **'Bloomfield Formation'** - Keighley, in press). The formations of this lowermost succession are classified as the Horton Group. Placement of contacts associated with the Albert Formation in particular has been a problem, since in both core and outcrop the succession contains intervals of interbedded red and grey strata that are of sandstone, shale, oil shale, and carbonate. These interbedded units represent interfingerings of laterally discontinuous lithotypes (beds commonly grade into one another), with successive occurrences of a particular lithotype having different spatial extent. Therefore, in any two vertical sections (outcrop or borehole), the first and last occurrences of a particular lithotype are unlikely to identify time horizons, or to delineate a similar thickness of intervening strata.

Historically, the base of the Albert Formation, when conformable with the underlying Memramcook Formation, was taken going downsection at "...where the red-beds become predominant, or where the lithological character changes sharply from fine grained to conglomeratic..." (Greiner, 1962, p. 220). A more precise definition was provided by St. Peter (1993) who defined the contact as being the first occurrence going upsection of grey strata containing wave-ripples, algal, ostracod, or fish fossils, or kerogenous lamination, above the basal red-bed conglomerate interval. Greiner's (1962) Frederick Brook Member, comprising oil shale and rare limestone and siltstone, was considered medial to his other two members. The upper and lower contacts were both considered gradational and, unfortunately, were arbitrarily placed, both upsection and downsection "...where the first shales low in bitumen and lacking in fish remains appear..." or alternatively "...where obviously different rocks become important, in general where sandstone and siltstone predominate." (Greiner, 1962, p. 229, p. 230). As pointed out by Carter and Pickerill (1985a), a further problem with Greiner's (1962) tripartite division is that the Dawson Settlement Member contains quite similar lithotypes to the Hiram Brook Member, making distinction between the two difficult unless the intervening Frederick Brook Member is present. There are additional problems in this subdivision of the Formation. In the north of the basin, Greiner (1962) describes the Formation as mostly of calcareous shale, argillaceous dolomite, and limestone, and lacking in oil shale, yet the succession is discussed within his commentary on the Frederick Brook Member. Also, where the Frederick Brook Member is absent, no guideline has been established as to whether the laterally equivalent grey beds should be assigned to the Dawson Settlement Member or the Hiram Brook Member. Keighley (in press) will be redefining the Frederick Brook Formation that clarifies the basal contact as the base of the first oil shale encountered in any given area. This is because no oil shale is described by Greiner (1962) for the type section of the Dawson Settlement Member. However, the contact between the two formations is unlikely to be a timeline because successive interfingerings of oil shale may have different lateral extent: the same interfingering of oil shale need not be the first encountered in different vertical sections.

Unfortunately, oil shale does interbed with calcareous shale, siltstone, sandstone, and thin limestone in what is described as the type section of the Hiram Brook Member by Greiner (1962). In order to better delineate this latter unit from the Frederick Brook

Member Keighley (in press) will revise the top of the Frederick Brook Member so that it corresponds to the top of the uppermost shale bed containing a distinct oil shale. In the Sussex area, this revision might also help distinguish characteristics of the Dawson Settlement Member and the Hiram Brook Member. The latter unit appears to lack coarse-grained to pebbly feldspathic sandstone but to contain a predominance of coarsening upward units.

The second succession is bounded from the first by a regionally significant angular unconformity that can overstep the Horton Group and lie on basement (St. Peter, 1993). The unconformity is considered related to the first structural inversion event in the Tournasian (Wilson, 2005). This red-bed succession consists of basal conglomerate (**Round Hill Formation**) transitional upward into shale and sandstone (**Weldon Formation**). Very locally, boreholes have intersected an evaporite succession (**Gautreau Formation**) at the base of the succession. In the Moncton area, these units and a tuff bed (the Boyd Creek Tuff) were all, until recently, thought part of the Horton Group. However, work by St. Peter (pers. comm, 2004) has correlated these units with the **Mill Brook Formation** conglomerate and **Dutch Valley Formation** mudstone in the Sussex area that are assigned to the Sussex Group.

Another regional unconformity, corresponding to the second inversion event of Wilson (2005) and thrusting event of Park et al. (2005), marks the boundary between the second and third tectonostratigraphic successions. The third succession again contains a red-bed conglomerate at its base, namely the poorly defined **Hillsborough Formation**. Conformably overlying this red-bed unit are the only definitive marine strata in the Upper Palaeozoic of the region. Several formations are defined, some of which are lateral equivalents. For instance, **Parleeville Formation** bioclastic wackestones and boundstones are considered to pass laterally into the laminated wackestones of the **'Devine Corner Formation'**, and interfingering and overlying these limestones are the sulfates of the **Upperton Formation**. In the subsurface of the Sussex region, the Upperton Formation passes up into evaporite of the **Cassidy Lake Formation**, which includes commercially mined potash units, and the overlying **Clover Hill Formation**, which consists of a lower anhydrite and an upper rocksalt. Towards the border with Nova Scotia, where potash is absent, interbedded limestone, conglomerate, and sulfate is assigned to the **Lime-kiln Brook Formation**, and is overlain by rocksalt of the **Pugwash Mines Formation**. These formations are collectively assigned to the Windsor Group, which is defined as a Viséan carbonate-evaporite-red-bed succession, and whose base is taken at the first carbonate with a marine fauna. Although biostratigraphic data all indicate a "lower Windsor subzone A" for New Brunswick strata, the basal units are sufficiently different lithologically that the basal boundstones should not be equated with the Gays River Formation in Nova Scotia, nor the wackestones with the Macumber Formation. Furthermore, it is likely that at least one significant unconformity is present within the Windsor Group of New Brunswick, and carbonate clasts in the Hillsborough Formation (Park, pers. comm. 2004) may be of Windsor origin. This would suggest erosion of the true temporal equivalents of the Gays River and Macumber formations, and inclusion of the Hillsborough Formation in the Windsor Group (cf. St. Peter, 1993).

Strata overlying the Windsor Group (variably in conformable, unconformable, and subsurface dissolution contact) usually consist of a basal mudstone (grey, carbonate cemented, passing up into red, variably carbonate cemented) overlain by red (and grey) sandstone or red conglomerate. The mudstone is of highly variable thickness, and near the basin margins can be rapidly displaced in the succession by conglomerate. In the Sussex area, the mudstone, sandstone, and conglomerate-dominated successions are termed the Poodiac, Scoodic Brook, and Wannamaker formations, though none of these terms has been adequately introduced. Similarly, the conglomerates around Hopewell Cape are yet to be properly introduced ('**Hopewell Cape Formation'** of St. Peter, in press), although Norman (1941) defined the **Maringouin** (red mudstone), **Shepody** (grey sandstone and mudstone), and **Enragé** (red-bed) formations in the surrounding area. Originally, these three formations were called the Hopewell Group (Norman, 1941), although general consensus is now to use the Mabou Group for this succession (although Hopewell Group has nomenclatural priority - Keighley, in press). Spore data indicates that the rocks are of late Viséan to Namurian age.

The fourth succession is deposited on the unconformity that marks the end of the Mississippian deposition (manifest as folding and thrust faulting in the Moncton Basin). It contains predominantly grey-bed sandstone and mudstone of the **Boss Point Formation** that is transitional upsection into the red-bed sandstone and mudstone of the **Grand Anse Formation**. These rocks, of Bashkerian, or ?latest Namurian to middle Westphalian A age (see discussion in St. Peter, 1993), lie with distinct angular unconformity on older Carboniferous strata and are included together in the Cumberland Group.

A fifth unconformity-bounded succession contains red-bed strata of the **Minto Formation, Salisbury Formation, Richibucto Formation, and Tormentine Formation**, collectively assigned to the Pictou Group, and which are dated as late Bashkirian to early Moscovian (Westphalian C to Stephanian).

2.3. Horton Group 'Lacustrine' Lithofacies

Throughout the Maritimes Basin Complex, the depositional setting of grey strata assigned to the Horton Group has long been attributed to lacustrine sedimentation (e.g. Bell, 1929; Greiner, 1974; Hesse and Reading, 1978; Carter and Pickerill, 1985b; Martel and Gibling, 1991). This reasoning was due to the abundance of fossilized plant material, kaolinite, analcime, and palaeosols, and the absence from these rocks of distinctly marine (e.g. tidal) sedimentary structures, normal marine invertebrates and, particularly, the distinct absence of any type 2 (marine phytoplankton and zooplankton) kerogens in the Albert Formation oil shale (e.g. Macauley and Ball, 1982). However, palaeontological questions remain. Greiner (1977) described crossopterygian fish (relatives of the marine coelacanth) from the Albert Formation near Moncton. Tibert and Scott (1999) recently suggested that ostracodes and agglutinated foraminifera in the Horton Bluff Formation near Wolfville, NS, are indicators of marginal marine brackish bay settings. R. Ryan (pers. comm. 2004) has also identified a trilobite from the formation.

The obvious way to reconcile the apparent paradox is to consider the sedimentary basins to be close to sea level (after all, the Windsor sea subsequently spread across the region). The outlet from the Horton Group lakes may have been into the ocean during periodic marine high stands and/or when the lake was not closed (e.g. as currently with Lake Maracaibo in Venezuela). Indeed, Higgs (2004) suggests the same situation for Namurian strata in SW. England and W. Ireland (highly controversially in DK's view). Significantly more research must be undertaken to address the problem.

Recent studies of the depositional settings in the Moncton Basin were by Carter and Pickerill (1985b), Pickerill et al. (1985) Smith and Gibling (1987), and St Peter (1992). Four settings are broadly recognised with complex spatial and temporal intergradation.

(1) alluvial fan-piedmont (or fan delta): Poorly sorted, mostly grey-green conglomerate and sandstone of variable particle size are thickest close to the basin margin with the Caledonia. Beds thin basinwards, interfingering with siltstone and shale containing caliche and mud-cracks. Structureless and matrix-dominated conglomerate at the margin likely represents debris-flow deposits. Inverse grading is present distally. Pebbly sandstone is often cross-stratified indicating aqueous

deposition, but sheet or channelized deposition is hard to determine.
(2) fluvial-floodplain-deltaic: Primarily interbedded, channelized, normally graded, pebbly sandstone to siltstone containing planar cross-strata, asymmetrical and symmetrical ripples, parting lineation, and conglomerate lags and lenses scour into, and are overlain by finely laminated, disrupted or massive shale, mudstone, siltstone, argillaceous dolomite and fine-grained sheet sandstone with climbing ripples and symmetric ripples. Root traces and cracks are common in the latter, floodplain facies. Delta-front sandstone deposits may also be present.
(3) marginal lacustrine (clastic shoreface, algal swamp, and carbonate mudflat): The shoreface lithofacies comprise 0.5 - 1.0 m thick fine- to medium-grained, well sorted, locally channelized sandstone (with desiccation and/or syneresis cracks, rain prints, symmetrical ripples, flaser beds, climbing ripples, horizontal lamination, low-angle cross-strata, and convolute lamination) interbedded with 0.2 - 0.4 m thick, laminated or soft-sediment deformed shale or mudstone. Pisolitic beds and laminated carbonate mats (brecciated) reflect algae-related deposits. The carbonate mudflat lithofacies consists of thinly interbedded dolomitic shale, siltstone, kerogenous mudstone, and lenticular very-fine-grained sandstone that contain plant and fish debris, desiccation cracks, rare rain prints, and carbonate nodules.
(4) lacustrine (perennial - 'deep'): Rhythmically bedded 'oil shale' (detailed below) with pyrite, lack of bioturbation, and delicate complete vertebrate fossils accumulated at hypolimnion depths (typically anything below ~ 15 m water depth). Thin sandstone interbeds represent density flow deposits.

A general distribution has two major delta systems (the Norton and Stoney Creek deltas) in the southeast and southwest, separated by fans on the southern Caledonia margin, mudflats on the north, and the lake in-between. This lateral facies heterogeneity indicates the presence of an asymmetric basin, with an active Caledonia footwall and "...a passive 'hinge' margin in the north." (Carter and Pickerill, 1985, p. 71). A general overstepping from the south and southwest was identified, accompanied by an accompanying climatic evolution that initially resulted in an open basin, but later a closed evaporitic basin. This model has proven to be quite robust, although recent work would now put the evaporite of the Gautreau Formation in a later tectonostratigraphic cycle. A predominantly closed basin need not be implied for the later stage of infilling.

2.4. Petroleum

Petroleum (from the Greek for "rock oil") is a term with varied use. Following the on-line glossary of Natural Resources Canada, it describes gaseous, liquid, or solid hydrocarbon compounds, which may be variably admixed with other chemical compounds and elements, and present within or extracted from the earth.

In New Brunswick, the petroleum industry began in 1849 when a solid vein of bitumen (Albertite) was discovered transecting high grade oil shales along Fredericks Brook at Albert Mines (St. Peter, 1988). This vein was mined for around 20 years. The material was shipped to Boston and Philadelphia, where it was distilled for the production of lamp oil (kerosene), in place of whale oil (the petroleum industry saved the whales!). Circa 1859, a Pittsburgh refiner drilled 3 wells near the village of Dover, on the east side of the Petitcodiac River (southeast of Moncton), producing small quantities of oil (Howie, 1968). Similar quantities were produced from other wells in the vicinity between 1876 and 1879, and 1903 to 1905 (St. Peter, 1988).

In July 1909, Maritime Oilfields Ltd. (MOL) of London, England, discovered the Stoney Creek field on the other side of the Petitcodiac River, but on strike with the Dover field. The discovery well (MOL #4) flowed 500 mcf gas per day and several barrels of oil, and MOL #5 had a flow of 1 mmcf per day. Hillsborough and Moncton were piped natural gas from 1912 until the 1970s. 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 until its closure in 1991. Total oil in place is calculated at ~2.1 x 106 cubic metres with <5% primary recovery. Natural gas composition is methane (~75%), ethane (~20%), non-hydrocarbons (5%), with undetectable amounts of sulfur (Howie, 1968). Crude oil composition is paraffins (~70%), Olefins (~5%), naphthenes (~20%), aromatics (~5%), with an API gravity of ~35 at 60^{o}F (St. Peter, unpublished data) - a composition very typical of type 1 kerogens of lacustrine origin. These Albert Formation (Fredericks Brook Member) reservoirs are in sandstone lenses up to 30 m thick (Howie, 1968) at depths ranging mostly from 600m to 900m. They are interbedded and overlain by kerogenous mudstone and oil shale (the source rock) in a stratigraphic trap (gently south-dipping sandstone pinches out up-dip into mudstones - Howie, 1968). Cleaner Albert Formation sandstone has porosities of 8 - 20% and permeabilities of up to 100 milliDarcies (Chowdhury and Noble, 1992).

During the 1990's the construction of the Maritimes and Northeast Pipeline (M&NP) to transport offshore (Sable Island) natural gas to the Boston area, sparked renewed petroleum exploration in the region. A natural gas discovery (the Downey prospect - yet to be fully evaluated) was made 5km south of the Stoney Creek field in 1998 by MariCo Oil and Gas Corp.

Of greatest note is the summer 2000 Corridor Resources Inc./Potash Corporation of Saskatchewan McCully #1 discovery well east of Sussex, from which a 22 m sandstone package at around 2.3km depth (considered lower Hiram Brook Member of the Albert Formation) flowed at 2.5 mcf per day. The field as a whole covers over 7000 acres and is a large anticlinal structure with simple four way closure. Interformational faults, stratigraphic pinch outs, and the base Sussex Group unconformity are likely also factors in trapping the gas. To date, drilling has been concentrated on the northwest limb, and no wells are considered to have penetrated below the upper Fredericks Brook Member. The gas pressure is consistently over 500 psi above the hydrostatic pressure at this depth, indicating a >500 m gas column. However, porosities (average 8%) and permeabilities are low, because the sandstone is typically fine grained, and the small pore throats result in high capillary pressure. Residual bitumen can clog pore throats. Low thermal gradients in the basin, coupled with the overpressures can lead to hydrate formation. Furthermore, the sandstone is desiccated: in having very low water saturation (10-20%), irreducible water (Sw) capacity is much higher, meaning that any added water produces phase blocking.

The field is now in production, but at the moment supplying only power to the adjacent potash mill. The ongoing drilling program is aimed at increasing confirmed reserves in order to warrant construction of a 35 km lateral feeder to the M&NP. To date, two production wells have produced over 1.1 bcf of gas at an average of 2 million cubic feet per day (limited by demand) over 18 months. An independent report puts the proven and probable reserves, in just the northwest flank of the field, at 119 bcf.

There have been numerous attempts to commercially mine the oil shale of the Albert Formation. In 1927 several small adits were opened in the Rosevale area and production facilities constructed, but the Maritime Eduction Company went bankrupt before production began. The Atlantic Richfield Company (1968-9) and later Canadian Occidental Petroleum (1974 - 1996), AA (NB) Inc (1996-99), and Shell Canada Ltd (1998-1999) have drilled or trenched the oil shales in the Dover, Albert Mines, and Rosevale areas, but no production has been attempted. Macauley and Ball (1982) calculated that in the Albert Mines area, there is an in-place resource of 67 x 10^6 barrels of shale oil averaging 93.5 L of oil per tonne to a depth of 600 metres.

3) Structure, evolution, and lithostratigraphy of the Fundy Rift System

3.1. Introduction

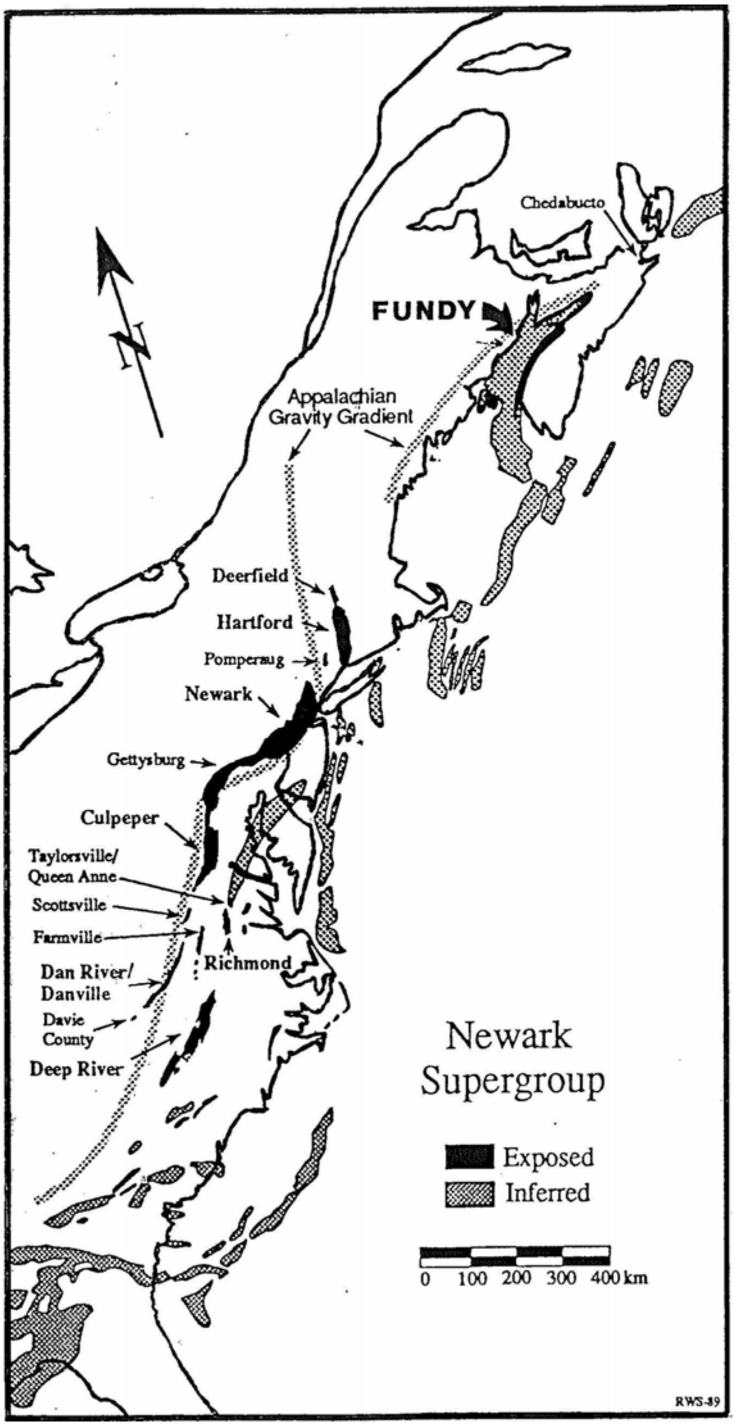


Figure 11 Distribution of Newark Supergroup basins, eastern North America (slightly modified after Olsen et al., 1989). The Fundy Rift is the largest of all Newark-type basins and covers an area of about 14,000 sq. km.

The basins of the Fundy Rift System represent the largest (14,000 sq. km), northernmost exposed remnant of Triassic-Jurassic synrift sedimentation along the east coast of North America (Figure 11).

These basins formed during the early synrift period of Atlantic rifting starting in the Middle Triassic. Their stratigraphic successions reflect underfilled lacustrine basin sedimentation in response to the dominantly extensional tectonic regime. The red beds in the Bay of Fundy region are genetically related to the Newark Supergroup of eastern North

America (Klein, 1962; Cornet, 1977; Van Houten, 1977; Froelich and Olsen, 1984). Nine major rift basins are preserved from South Carolina to Nova Scotia, along with numerous smaller basins and sub-basins, and many more buried beneath the more recent Atlantic Coastal Plain sediments and the continental shelf.

These rift basins were initiated during the opening phases of the proto-Atlantic Ocean by the break-up of the supercontinent Pangea, and closely followed the trend of the older compressional Appalachian Orogen (Manspeizer, 1988). Seismic and other data show these basins to be more complex than previously believed and that they owe their development to the extensional reactivation of pre-existing thrust faults and ramps formed during the Variscan Orogeny (Manspeizer, 1981; Peterson et al., 1984; Taylor and Resseter, 1985; Swanson, 1986). Most of these rift basins are half grabens, though some such as the Fundy, Minas, and Chignecto basins have been variously influenced by major transpressive faults. Each basin exhibits a trend in which its long axis is parallel to the main bounding fault, into which thousands of metres of continental sediments and basic volcanic rocks were deposited over a period of about 40-45 Ma (Olsen, 1986).

3.2. Regional Setting

Williams (1979) divided the northern Appalachians into several zones or accretionary terranes, which are attributed to the final closing of the late Precambrian-early Palaeozoic Iapetus Ocean (Figure 8). These exotic terranes (Williams and Hatcher, 1982; 1983) are separated from distinctly different adjacent terranes by major faults. The metallogenesis, metamorphism, stratigraphy, sedimentology, plutonism and tectonism of each terrane are distinctive.

In the Fundy region, two terranes are recognized which are termed the Avalon terrane and the Meguma terrane. The Avalon terrane is a composite of various accreted terranes characterized by late Precambrian volcanic and sedimentary rocks intruded by granites and overlain unconformably by shallow-water sediments of Cambro-Ordovician age. In New Brunswick, Avalon terrane rocks comprise most of the Fundy coastline with the largest single assemblage represented by the Caledonian Highlands. The Cobequid Highlands, bordering the north shore of Minas Basin, also consist of Avalon basement.

The Meguma terrane is the most eastern of all Appalachian terranes and consists of a sequence of Cambro-Ordovician slate and quartzite that are intruded by Devonian granite (Schenk, 1978). The Meguma is the largest terrane in this region, making up the entire southern half of Nova Scotia. Both the Avalon and Meguma Zones have undergone several orogenic events during their history.

The Minas Geofracture ("MGF", = Cobequid-Chedabucto fault zone), which is the most important structural feature in the Fundy region, separates the Avalon and Meguma terranes. It is a deep-crustal transverse fault that strikes east-west (Keppie, 1982; Mosher and Rast, 1984), with two phases of right-lateral movement proposed by Keppie (ibid): 475 km during the initial compressive phase of the Acadian Orogeny (early Middle Devonian) as well as normal motion (down to the north), and 165 km during the Pennsylvanian to Permian compressive Variscan (Hercynian-Alleghanian) Orogeny. The next phase of activity occurred during Middle Triassic to Late Jurassic, with about 75 km of left-lateral motion along the fault, and a compressive right-lateral phase in the Early Jurassic or Early Cretaceous with inversion of large structures within the Fundy Basin (Withjack et al., 1995, 1998). Minor movement along the MGF probably continued beyond the Early Cretaceous and perhaps even into the Cenozoic (King and MacLean, 1970).

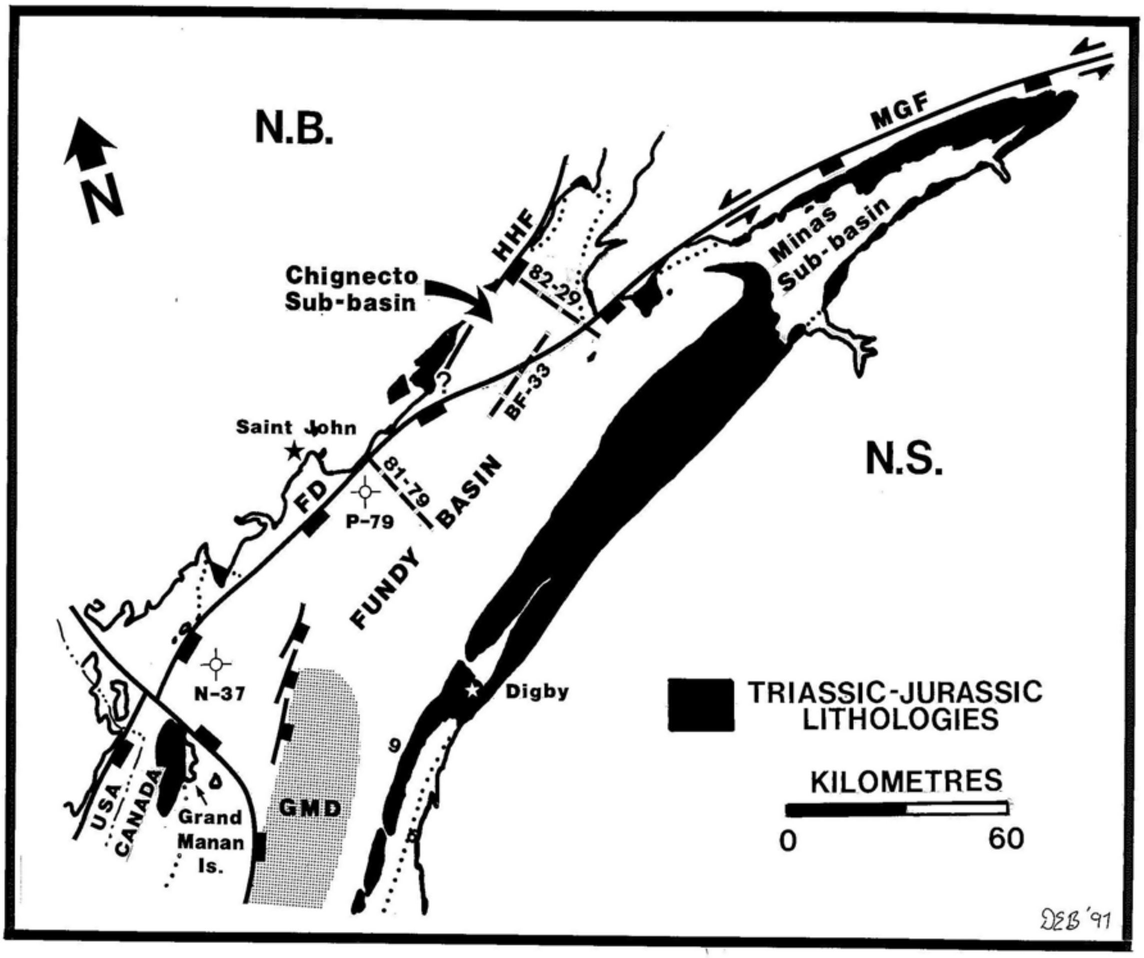


Figure 12: Distribution of Triassic-Jurassic rocks of the Fundy Group, Bay of Fundy area. The traces of the Minas Geofracture, Fundy Decollement and Harvey-Hopewell Fault are identified as MGF, FD and HHF respectively.

Early Mesozoic rocks are found in the Bay of Fundy region of Nova Scotia and New Brunswick (Figure 12). Their distribution is limited in New Brunswick to small, isolated fault-bounded grabens which defy correlation (Klein, 1962; Nadon, 1981; Nadon and Middleton, 1984, 1985). In Nova Scotia however, rocks of the Fundy Group (Klein, ibid) crop out along the entire Bay of Fundy coastline and along both shores of Minas Basin. Fundy Group lithologies are present offshore extending from Nova Scotia to within a few kilometers of the New Brunswick coastline and further out into the Gulf of Maine (Tagg and Uchupi, 1966; Ballard and Uchupi, 1972, 1975; Wade et al., 1996).

3.3. Age of the Fundy Rift System

The early Mesozoic rocks in the Fundy Rift System appear to range from early Middle Triassic (Anisian) to earliest Jurassic (Sinemurian) age (Figure 13 and Table 2). Age determination of the sediments comes from studies of pollen and spore assemblages (Cornet, 1977; Dunay, 1975; Bujak, 1979; and Chevron, 1986), invertebrate fossils (Cameron, 1986), fish remains (Olsen et al., 1982; Olsen, 1988), and terrestrial vertebrate bones and footprints (Olsen and Baird, 1982; Olsen et al., 1987), while the ages of the volcanics are derived from K/Ar dating (Hodych and Hayatsu, 1988). The stratigraphic sections in Nova Scotia are believed to be younger than those in the New Brunswick grabens, although paucity and poor preservation of biostratigraphic material in these grabens applies constraints on the accuracy of their dating.

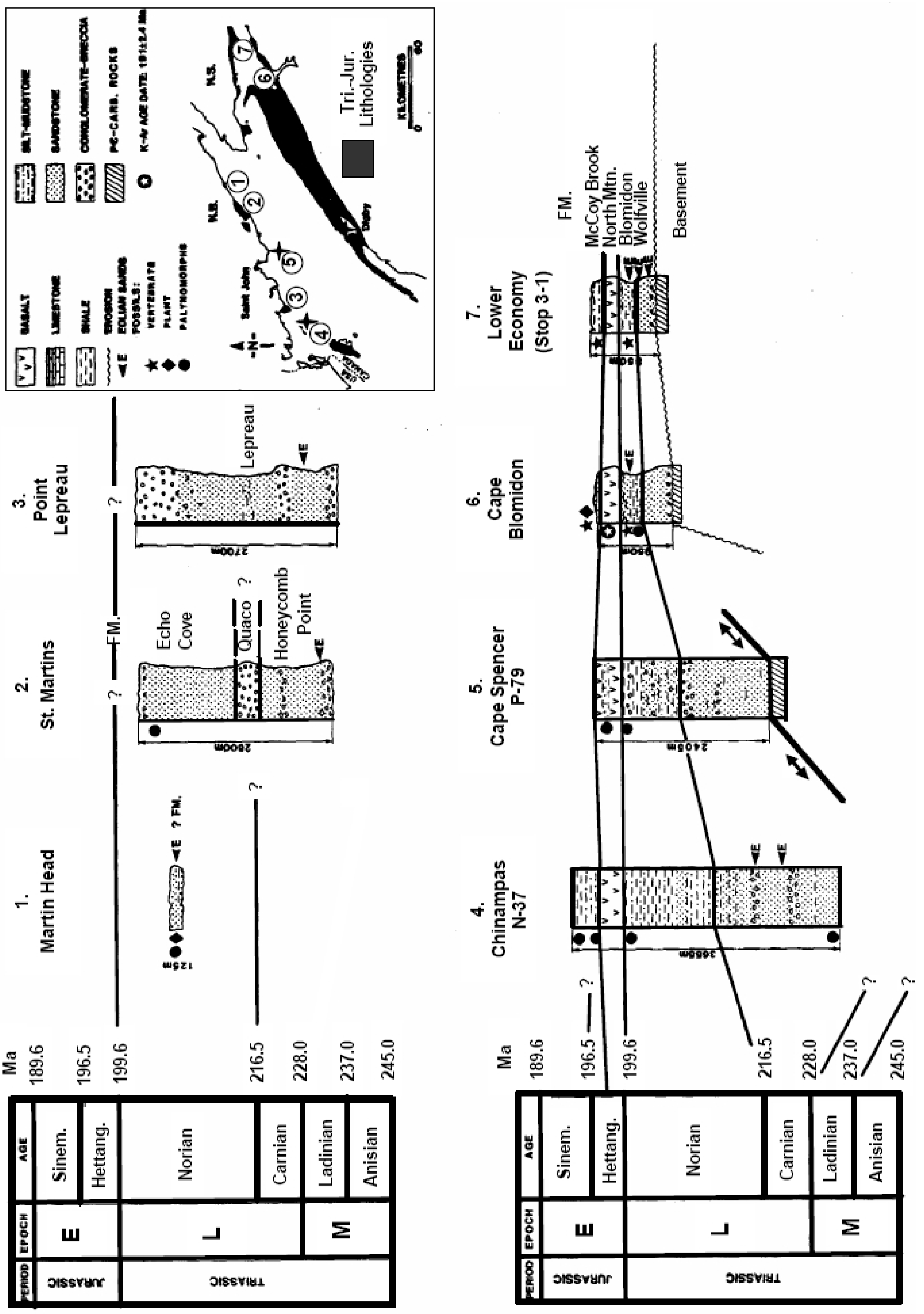


Figure 13: General stratigraphic relationships and ages between selected outcrop sections and offshore wells, Fundy Rift System. Offshore thicknesses from well data unless otherwise indicated. Time scale from Gradstein et al. (2004).

Both offshore wells, although located just off the New Brunswick coast, mirror the sections present in Nova Scotia in both age and stratigraphy. Palynomorphs from the Chinampas N-37 well identifies the Triassic-Jurassic boundary about 740 m below the basalt (Dunay, 1975). A similar study by Bujak (1979) however, places this boundary about 125 m above the basalt, and assigns a Hettangian to Sinemurian age to lithologies above this point. The palynostratigraphy of the Cape Spencer P-79 well though, is similar to that described by Dunay (ibid) for N-37, as Hettangian-aged strata extend about 490 m below the basalt (Chevron, 1986). The presence of Jurassic palynomorphs well below the basalt in these wells is at variance with the established onshore Triassic-Jurassic boundary which is located immediately at the base of the basalt. Perhaps this represents a difference in subsidence rates between sediments along the basin margin, versus those adjacent to the main bounding fault and within the depocentre just east of Grand Manan Island (Brown, 1986). A more probable interpretation would be that the sampling procedures during the drilling of these wells was deficient and therefore compromised the resolution of their palynostratigraphy. The Triassic-Jurassic boundary is probably at most a few tens of metres below the basalt.

New Brunswick Sections (1 – 3)					
FORMATION	THICKNESS (m)		GENERAL LITHOLOGIES	DEPOSITIONAL FACIES	REFERENCES
	Onshore (outcrop)	Offshore (seismic)			
Echo Cove	850-1300	?	redbed sandstone, conglomerate, breccia, minor siltstone	alluvial fan, fluvial; braidplain	Nadon & Middleton, 1984
Quaco	190-300	?	conglomerate, sandstone	fluvial braidplain	Nadon & Middleton, 1984
Lepreau	2700	?	redbed conglomerate and minor sandstone	proximal alluvial fan, minor fluvial and aeolian dune	Nadon & Middleton, 1984
Nova Scotia & Offshore Well Sections (4 – 7)					
McCoy Brook	200	357 (3000+)	redbed conglomerate, sandstone, siltstone, shale	alluvial fan, aeolian, fluvial, playa, lacustrine	Hubert & Mertz (1984, Olsen et al. (1989), Tanner & Hubert (1991; 1992), Wade et al. (1996)
Scots Bay	8	? (?)	limestone, shale, minor sandstone, chert	shallow lacustrine	Birney (1985), De wet & Hubert, (1989)
North Mountain	260	333 (~1000)	tholeiitic basalt	continental fissure-type volcanic flows	Stevens (1987), Olsen & Schlische (1990), Wade et al. (1996)
Blomidon	400	1168 (~2500)	redbed siltstone, mudstone, shale and minor sandstone	playa, sandflat, shallow to deep lacustrine, aeolian	Klein (1962), Hubert & Hyde (1982), Mertz & Hubert (1990), Olsen & Schlische (1990), Wade et al. (1996)
Wolfville	400	1708 (3000+)	Redbed sandstone and conglomerate and minor siltstone and shale	alluvial fan, fluvial braidplain, aeolian, deepwater lacustrine	Klein (1962), Fraley (1982), Hubert & Forlenza (1988), Wade et al. (1996)

Table 2. Table of formations and associated lithologies of the Fundy Group, Fundy Rift System.

In summary, the age range of the lower Mesozoic sediments in the Fundy Rift System appears to extend from the Anisian to at least Sinemurian age. Olsen (1989) suspects that the age of the McCoy Brook Formation may extend into the Pliensbachian and Wade et al. (1996) believe it is an even Aalenian age. This age range represents the most extensive and continuous stratigraphic succession in the entire Newark Supergroup.

3.4. Evolution of the Fundy Rift System

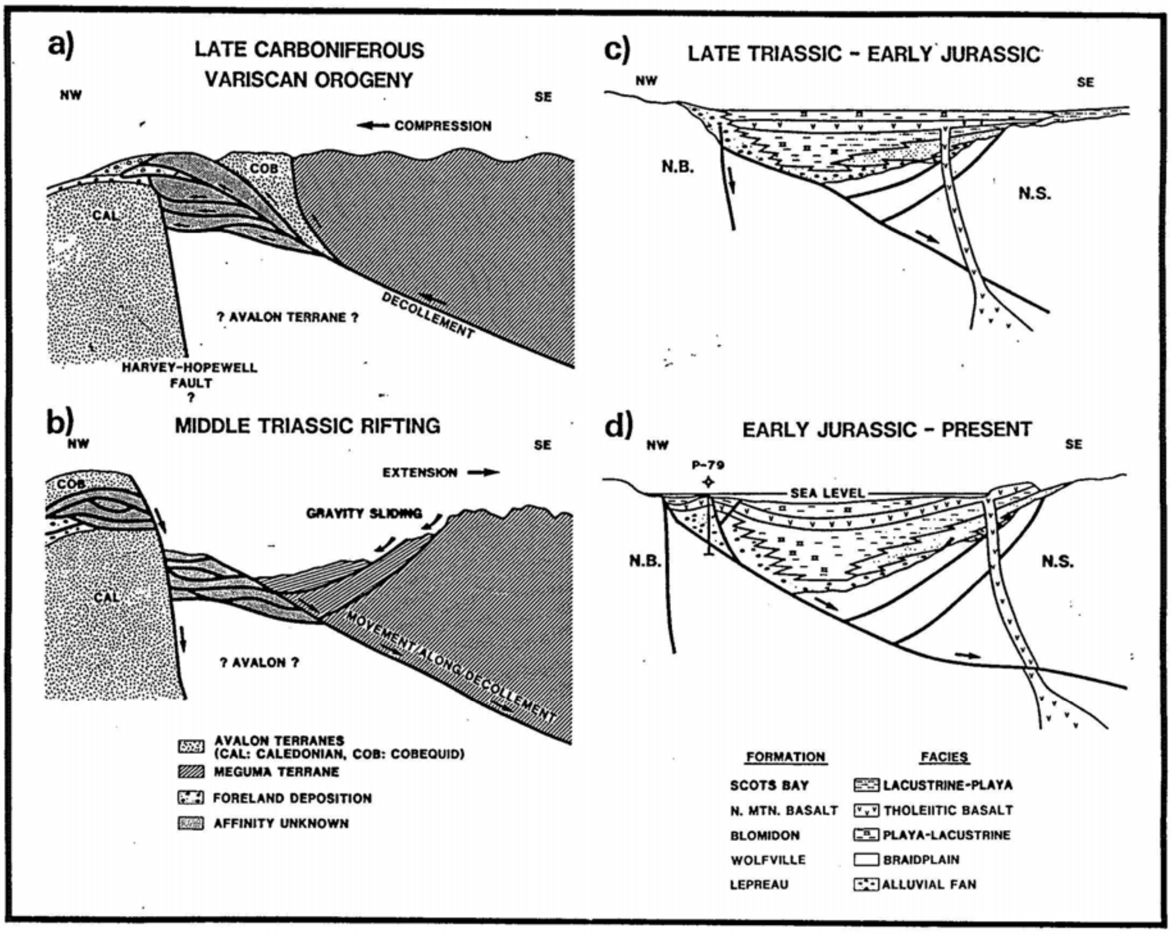


Figure 14. Evolution of the Fundy Basin; see text for details.

3.4.1. Late Devonian - Late Carboniferous (Figure 14a)

The Variscan Orogeny in the Bay of Fundy region represents the final collision between the Meguma and Avalon terranes, with the westward dextral motion of the Meguma terrane along the Minas Geofracture (MGF) being transmitted into a northwest-directed thrust component. Late Carboniferous (Westphalian B) sedimentary rocks along the south coast of New Brunswick record this Variscan thrusting (Plint and van de Poll, 1982, 1984; Caudill and Nance, 1986). Seismic profiles across the Bay of Fundy reveal a shallow (10-15°) southeast dipping ramp or decollement beneath the Triassic-Jurassic section that appears to flatten and extend beneath the Meguma terrane. This feature, the Fundy Decollement (Wade et al., 1996; Brown, 1986), is observed in seismic profiles extending from just east of the Chinampas P-79 well to the eastern shore of Grand Manan Island, and extends beneath the Bay of Fundy and beyond the limit of industry seismic resolution (6 seconds two-way time or approximately 12 km). Deep crustal seismic profiles confirm that this feature extends down to at least this depth and probably flattens out at the top of the Moho beneath the Meguma terrane at about 10 seconds two-way travel time (Keen et al., 1991). Beneath the decollement are thin (1-2 km thick) stacked intervals which dip at an angle of 5-10° to the southeast. These are interpreted as northwest-directed thrust sheets of unknown affinity, possibly sediments of early Carboniferous age (Wade et al. ibid. Brown, ibid), or older lithologies from an Avalon-type terrane (Keen et al., ibid).

3.4.2. Late Carboniferous - Middle Triassic (Figure 14b)

Fluvial siliciclastics of the Pictou Group and its equivalents blanketed most of the Maritime region, (Westphalian C to Early Permian; Hacquebard, 1972), covering all Lower Carboniferous strata and in places overstepping basement complexes. These sediments are mildly deformed to undeformed, and mark the end of the Variscan Orogen. No Early Triassic sediments are recorded in the Fundy region, probably because of widespread erosion. Middle Triassic sediments of the Honeycomb Point formation are now believed to be of Permian age based on paleomagnetic data (Olsen et al. 2000, 2003).

3.4.3. Middle Triassic - Early Jurassic (Figure 14c)

The oldest Mesozoic sediments are those of the Wolfville Formation found at Carr's Brook/Lower Economy along the north shore of Minas Basin. An Anisian age is also suspected for some of the strata in the St. Martins area of New Brunswick (B. Cornet, pers. comm., 1987). It is believed that major crustal doming prior to rifting resulted in a northward-dipping palaeo-slope in the Fundy area, thus accounting for the northerly and easterly-directed palaeocurrents recorded in Triassic sediments of New Brunswick (Nadon and Middleton, 1984). Based on palaeomagnetic data, Olsen et al. (2000, 2003) indicate that Nadon and Middleton’s (ibid.) Honeycomb Point Formation is in fact of Permian age.

Following crustal doming, and during the deposition of the Middle Triassic sediments in New Brunswick and along the trace of the MGF in the Minas Sub-basin, tension on the Meguma terrane was relaxed (possibly due to cooling), which allowed the block to move eastward (sinistrally). Pre-existing structural lineaments facilitated this motion in the vertical sense (MGF) and sub-horizontal sense (Fundy Decollement). Compressional facies of the Variscan Orogeny were relaxed and superseded by extensional facies. The probable rapid back-sliding of the Meguma terrane down the decollement and along the MGF resulted in a large, deep basin (Fundy), and the smaller Minas and Chignecto sub-basins. The older, north-dipping Triassic (and Permian?) grabens in New Brunswick were abandoned, and sedimentation was now directed southeastwards toward the main Fundy Basin and to a lesser extent the Chignecto Sub-basin. Sedimentation in the Minas Sub-basin was influenced by both vertical and transverse movement along the MGF.

Coarse siliciclastic sediments shed across the border faults formed large alluvial fan complexes, while from the opposite side of the basins piedmont braidplain systems were established. Both systems deposited sediments into a coeval basinal lacustrine environment in the basinal sags. Evidence of synsedimentary tectonism is observed in sediments the margin of the Minas Subbasin. The MGF was expressed as a positive structure where it crossed the basin diagonally and produced thrusts, shatter zones, listric normal faults and local unconformities within the sediments. An uninterrupted volcanic event followed with the extrusion of tholeiitic basalt reaching a seismically-defined maximum thickness opposite Grand Manan Island of about 1000 m (Wade et al., 1996). Following this event, fluvial, aeolian and playa-lacustrine facies were re-established (McCoy Brook and Scots Bay formations).

3.4.4. Early Jurassic - Recent (Figure 14d)

Deposition of the nearly 12 km of sediments in the Fundy Basin probably ceased by the end of the Early Jurassic, though sedimentation in the Chignecto and Minas subbasins may have continued into the Middle Jurassic. Seismic data reveal that most of the large scale structural features mapped (anticlines, synclines and planar normal listric faults) incorporate all known Mesozoic strata. Cornet and Olsen (1985) and Traverse (1987) believe that the upper age limit of the McCoy Brook Formation may extend into the Pliensbachian. Based on thickness of the McCoy Brook in the Fundy Basin as defined by seismic, and comparison with rates from equivalent successions in the Newark Basin (Schlische and Olsen, 1990), Wade et al. (1996) calculate a minimum Aalenian age for this sequence. If correct, this would indicate that this phase of tectonic adjustment probably occurred in the early Middle Jurassic (Aalenian) or later. Some of the structural styles within the basin record compressional tectonism indicating dextral motion along the MGF sometime in the Middle Jurassic to Early Cretaceous (Withjack et al., 1995). Minor structural adjustments probably continued, but without major movement of the MGF. King and MacLean (1970) suggest some Cenozoic movement along the MGF. The present form of the Bay of Fundy was probably initiated by late Mesozoic subareal erosion and later modified by glacial action during the Pleistocene.

3.5. Formations and Facies Development

The stratigraphic relationships of onshore early Mesozoic rocks in the Fundy Rift System represent a stratigraphic succession of basin-filling and fining-upward sequence of continental sediments derived from the surrounding basement complexes (Klein, 1962) (Figure 13 and Table 2). The basin-margin grabens in New Brunswick contain conglomerate and sandstone of the Lepreau, Quaco and Echo Cove formations. Along the Nova Scotia margin, alluvial, fluvial and aeolian sandstone and conglomerate of the Wolfville Formation form the base of the succession. Playa, lacustrine and aeolian sediments of the Blomidon Formation disconformably overlie the Wolfville. The North Mountain Basalt, a tholeiitic sequence, conformably overlies the Blomidon and is revealed by seismic to extend over the entire Fundy Basin (Brown, 1986; Wade et al., 1996). Limestone, chert and shale of the Scots Bay Formation, and sandstone and siltstone of the equivalent McCoy Brook Formation, subsequently overly the basalt.

Manspeizer (1981) showed that generally, three sedimentary facies can be recognized in all Newark-type basins. These facies are present in Fundy Rift System:

(1). Adjacent to the border fault, coarse boulder conglomerate and sandstone were deposited in alluvial fan and fan delta complexes. This border fault facies (Lepreau, Quaco, Honeycomb Point, Echo Cove and McCoy Brook formations) interfingers basinward with lacustrine shale and silt.

(2). Proximal fluvial braidplain sandstone and conglomerate were deposited in a marginal piedmont facies opposite the border fault facies (Wolfville and McCoy Brook Formations).

(3). The central basin facies varies between playa and lacustrine environments. The playa environment is represented by red mudstone, siltstone and caliche soils of the Blomidon and McCoy Brook Formations, whereas limestone, shale and siltstone of the Scots Bay Formation are representative of the lacustrine environment.

Aeolian sandstone is recognized in strata of all facies except the lacustrine in the Fundy Rift System (cf. Hubert and Hyde, 1982; Hubert & Mertz, 1980, 1984; Hubert et al., 1983). Syndepositional volcanism and extensional tectonism are common. Basalt flows, dykes and sills are dated to approximate the Triassic-Jurassic boundary (North Mountain Formation).

A more detailed delineation of sedimentary/environmental facies is outlined by Olsen (1988). His subdivision is based largely upon the characteristic fossil assemblage representative of each environment, the categories being: (1) fault-scarp breccias and synsedimentary grabens; (2) alluvial fans and river systems; (3) swamps and marginal lakes; (4) deltas and shorelines; (5) open water, shallow lakes; and (6) deep-water lakes. Other environments such as aeolian dune fields and coal measures are basin-dependent, reflecting the basin's own unique structure, geographic position, etc. (Olsen, ibid).

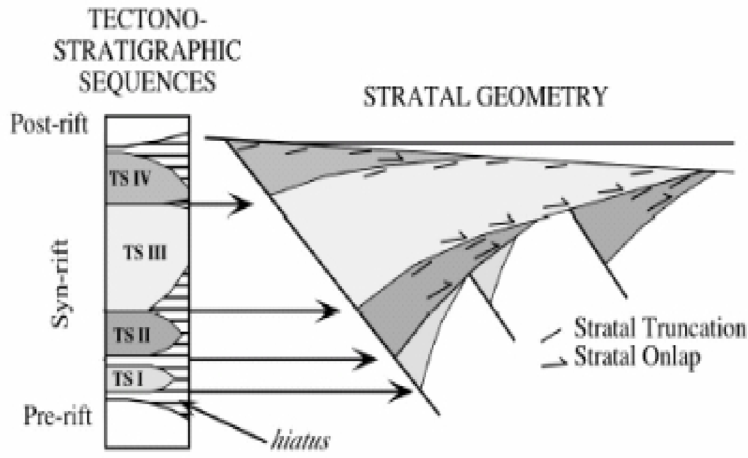


Figure 15: Idealized Atlantic early synrift basin showing unconformity-bounded tectonostratigraphic packages (Olsen, 1997). TS-I is eastern North America is very poorly exposed and so may be a prerift deposit and its wedge-shape profile is therefore speculative. The synrift TS-II sequence is areally more restricted and wedge-shaped than the overlying TS-III, and the TS-III to TS-IV transition is probably related to an increase in extension rates (increasing seafloor-spreading?) (Olsen, ibid.).

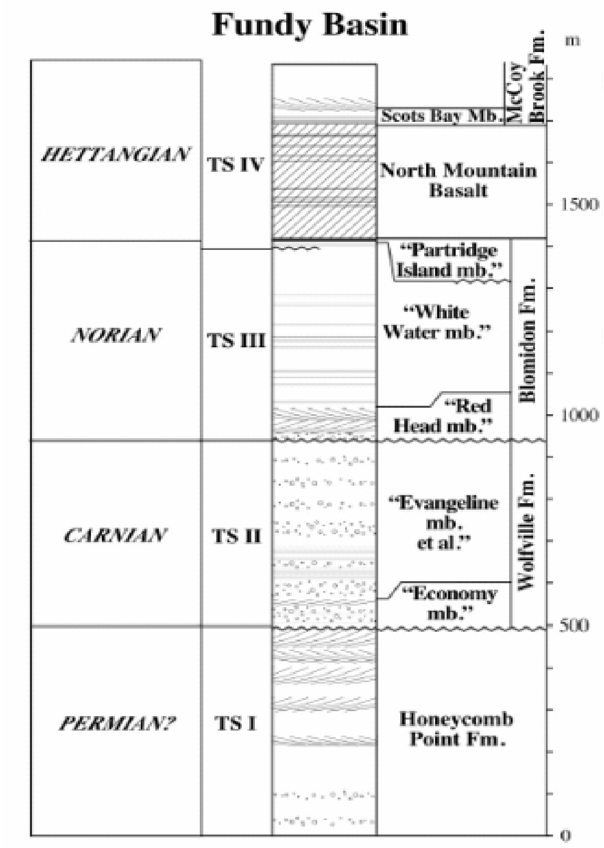


Figure 16: Tectonostratigraphy and age of the Fundy Basin (Olsen et al., 1997). Units noted as 'members' are informal terms as proposed by Olsen et al. (2000)

Olsen (1997a) has regrouped and later redefined (Olsen 1997b) the sedimentary successions within the rift basins of the Central Atlantic Margin of North America (Newark Supergroup) and Morocco in a tectonostratigraphic context. Four unconformity-bounded tectonostratigraphic sequences ("TS") are recognized: TS I – initial synrift fluvial and aeolian (Anisian); TS II – early synrift fluvial and alluvial (Ladinian to Carnian); TS III – middle synrift fluvial, playa and lacustrine (Carnian to Rhaetian); TS IV – late synrift lacustrine, playa, fluvial and volcanic (Hettangian to Pliensbachian (or Aalenian) (Figures 15, 16). Where drilled, equivalent basins offshore (Scotian Basin) appear to reflect these divisions although with greater quantities of evaporite deposits. These mostly continental half graben basins record sedimentation through time and attendant climatic change as the North American plate drifted northwards over an approximate 30 million year or greater time period.

3.6. The Minas Sub-basin

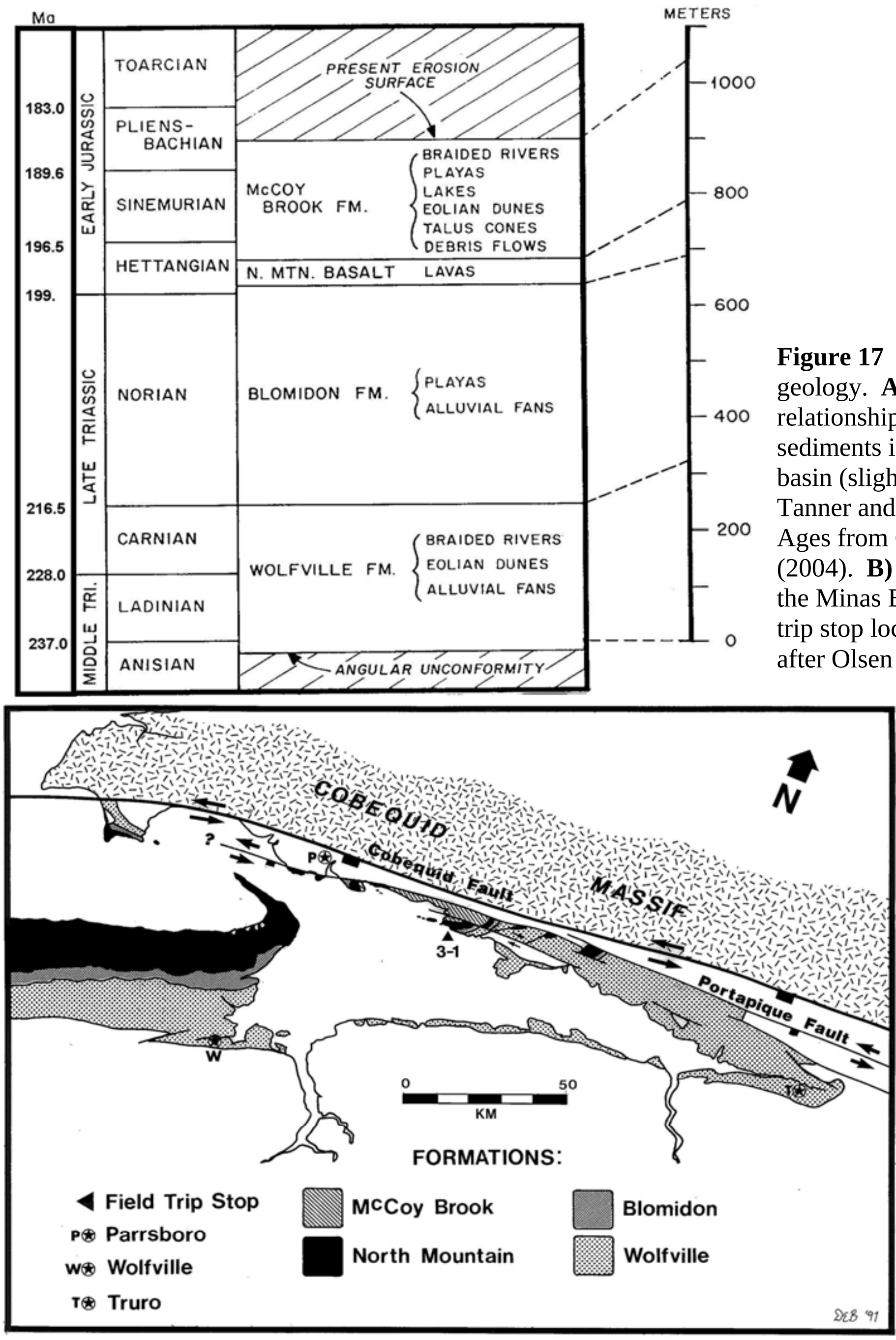


Figure 17 Minas Sub-basin geology. **A)** Stratigraphic relationships of Fundy Group sediments in the Minas Sub-basin (slightly modified after Tanner and Hubert, 1991). Ages from Gradstein et al. (2004). **B)** Geological map of the Minas Basin area, with field trip stop location (modified after Olsen et al., 1989)

The greatest outcrop exposure of the Fundy Group occurs on the margins of the Minas Sub-basin where over 1 km of sediments and basalt are exposed, primarily as sea cliffs (Figure 17). The Minas Sub-basin includes the area covered by the geographical Minas Basin, the Minas Channel and Cobequid Bay. The Sub-basin is bounded to the north by the faults of the Minas Fault Zone which merge westward with the boundary faults of the Fundy Sub-basin. The southern margin of the Sub-basin is presumed to be nonfaulted. The transtensional Minas Sub-basin is depositionally continuous with the Fundy subbasin, but differs in structural style from the deeper half-graben of the Fundy Sub-basin bounded to the west by southeast dipping normal faults (Figure 17B).

The **Wolfville Formation** is the basal unit in the Minas Subbasin and is comprised of reddish conglomerate, sandstone and shale deposited unconformably on pre-Triassic basement as alluvial fans and proximal to distal braided stream sheetflood deposits. On the north side of the Minas Sub-basin, the Formation crops out in fault-bounded blocks to the south of the Cobequid Fault in the area around Advocate Harbour to the west and is more or less continuously exposed to the south of the Portapique and Riversdale faults from Red Head to as far east as Valley Station, east of Truro. On the southern margin of the Sub-basin, exposure is nearly continuous on the coast between Valley Station and Paddy Island and the outcrop belt continues as far west as St. Mary’s Bay west of Digby.

The thickness of the Formation is approximately 360 m at the type section from Paddy Island to Kingsport (Klein, 1962), but thins to zero on the north shore of the Minas Basin, a result of synsedimentary fault-block topography (Olsen and Schlische, 1990). Thicker accumulations of sediment lie beneath the Bay of Fundy. Over 1700 m of sediments correlated as Middle to Late Triassic were drilled in the Mobil Gulf Chinampas well in the Fundy Sub-basin. Interpretation of seismic sections across the Bay of Fundy indicates that over 3 km of correlative section occurs in the deepest part of the Fundy Sub-basin (Wade et al., 1996).

The age of the Formation is assigned on the basis of a Late Carnian herpetofauna including saurians, synapsids, rhynchosaurids, rauisuchids, and labyrinthodont amphibians from middle Wolfville beds in Kings and Hants Counties and a Norian ichnofaunule (rhynchosauroid) from Upper Wolfville beds at the type location on Paddy Island (Baird and Olsen, 1983; Olsen and Baird, 1986; Olsen, 1988). A fault-bounded exposure of redbeds on the north shore of the Minas Basin in Colchester County containing osseus remains of labyrinthodont amphibians and procolophonid, synapsid and saurian reptiles has been dated as Anisian (Baird, 1986). This coastal section at Carrs Brook east of Economy Mountain, the TS-I "Lower Economy beds" of Olsen (1997), comprises steeply dipping (c.45^{o}) sandstone, conglomerate and mudstone of fluvial and aeolian origin.

The **Blomidon Formation** conformably overlies the Wolfville Formation except where it steps over basement on the north shore of the Minas Basin (Olsen and Schlische, 1990). Exposures on the north shore of the Minas Basin occur in fault blocks from Advocate Harbour to Economy Mountain and to the south from the eastern side of the Blomidon Peninsula westward to the shore of St. Mary’s Bay on the eastern side of the Fundy Sub-basin.

The Formation is roughly 315 m thick at the type section between Cape Blomidon and Paddy Island on the south shore of the Minas Basin. Syndepositional faulting causes the Formation to thin to as little as 10m on the north shore (Olsen and Schlische, 1990). Nearly 1200 m of Late Triassic strata correlative with the Blomidon Formation have been drilled in the Fundy subbasin in the Chinampas well and possibly as much as 3 km of correlative section occurs in the deepest area of the Fundy Sub-basin (Wade et al., 1996). It comprises reddish-brown to grey sandstone, siltstone and claystone deposited in lacustrine, playa, sandflat, and minor aeolian and fluvial environments (Hubert and Hyde, 1982; Mertz and Hubert, 1990). Cyclic alternations of coarse and fine facies have been interpreted as representing tectonic autocycles controlled by differential basin subsidence (Mertz and Hubert, ibid.) and Milankovitch-frequency climatic cycles (Olsen et al., 1989). The presence of gypsum, relict evaporite crusts and sand-patch fabrics indicate a period of aridity during deposition (Smoot and Olsen, 1988; Mertz and Hubert, 1990).

A Norian age for most of the Formation is assigned on the basis of a fresh water fauna consisting of crustacean (darwinulid and candonid ostracodes, and conchostracans) and fish remains (redfieldiid) and ichnotaxa (rynchosauroid and *Grallator* (Olsen, 1988; Olsen et al., 1989) as well as osseus remains (phytosaur) (Olsen et al., 1989). A Hettangian palynoflora dominated by *Corollina meyeriana* at the top of the Blomidon Formation indicates the position of the Triassic-Jurassic boundary within the upper 10 m (Cornet, 1977).

The overlying **North Mountain Formation** is exposed nearly continuously on the southern side of the Minas Sub-basin as cliffs or highlands for nearly 200 km southwest from the Blomidon Peninsula along the eastern margin of the Bay of Fundy. Fault-bounded exposures also occur as headlands on the north shore of the geographic Minas Basin between Cap d'Or and Economy Mountain.

The multiple flows of tholeiitic basalt attain a total thickness varying from 100 m on the north shore of the geographic Minas Basin to 270 m along the Bay of Fundy (Stevens, 1987; Olsen and Schlische, 1990). A maximum thickness of 333 m is recorded beneath the Bay of Fundy from well data (Olsen and Schlische, 1990). However, Wade et al. (1996), using seismic profiles and well velocity data claim the basalt is at least 1000 m thick in the depocentre southeast of Grand Manan Island. A Hettangian age for the Formation is indicated by its position between the Blomidon Formation and the overlying Hettangian Scots Bay and Hettangian-Pliensbachian McCoy Brook formations. This is in agreement with the most recent radiometric date of 202± ma (Hodych and Dunning, 1992).

Earliest Jurassic sedimentary rocks overlying the North Mountain Basalt attain a maximum thickness of over 350 m beneath the Bay of Fundy in the Chinampas N-37 well and may be over 2.5 km thick in the depocentre of the Fundy subbasin (Wade et al., 1995). Limited exposures of these rocks occur along the north (McCoy Brook Formation) and south (Scots Bay Formation) shores of the Minas Basin. Outcrops of the Scots Bay Formation occur only along the shore of Scots Bay on the west side of the Blomidon Peninsula in isolated synclinal basins. These basins have been interpreted as circular collapse structures in the North Mountain Basalt by Stevens (1987).

Exposures of the **Scots Bay Formation** lie directly on the surface of the North Mountain Basalt and range in thickness from 2 m to just over 9 m (De Wet and Hubert, 1989). The Formation comprises red and green calcareous siltstone, crudely bedded, commonly silicified limestone, thick-bedded chert, stromatolitic limestone, and brown, cross-bedded sandstone (De Wet and Hubert, ibid.; Olsen et al., 1989). The strata, containing charophyte debris, fish bones, ostracodes, gastropods, and conchostracans, represent deposition in a shallow, oxygenated, oligotrophic to eutrophic lake of shoreline to offshore sediments (Birney, 1985, Cameron, 1986). This lake may or may not have been widespread across the rift valley (Good et al., 1994). The cherts are attributed to hydrothermal vents on the lake floor (De Wet and Hubert, 1989). Despite the diverse fauna, no age-specific forms have been found. A Hettangian age for the Scots Bay Formation is inferred from its conformable position above the basalt.

Strata of the Lower Jurassic **McCoy Brook Formation** are exposed on the north side of the Minas Sub-basin between Clark Head and Old Wife Point south of the Portapique fault (Donahoe and Wallace, 1982). These rocks were initially assigned to the Blomidon Formation by Powers (1916) and Klein (1962). Liew (1976) first recognized the Jurassic age of outcrops on the north side of the basin and correlated them with the Scots Bay Formation, although the designation McCoy Brook Formation for these sediments both onshore and offshore is in wide usage today.

A Lower Jurassic age is supported by the stratigraphic relation to the underlying North Mountain Basalt of Hettangian age, ichnotaxa including therapod and fabrosaurid dinosaurs (Olsen, 1981; 1988), fish remains, and the osseus remains of synapsid, saurian and therapod reptiles (Olsen et al., 1987; Olsen, 1988). Palynology suggests an age range of Hettangian to Pliensbachian for these sediments (Traverse, 1987).

McCoy Brook strata comprise sandstone, mudstone, conglomerate, and breccia of fluvial, lacustrine, playa, sandflat, alluvial-fan, aeolian, debris-flow, and talus origin (Hubert and Mertz, 1984; Olsen et al., 1989; Tanner and Hubert, 1991; Tanner and Hubert; 1992) exposed in fault-bounded sections with a maximum thickness of ~230 m.

Oil shale test pit - Albert Mines (pre-overgrown days), photo courtesy Clint St. Peter/NB DNR

GUIDE: Day 1

Introduction to lacustrine sequence stratigraphy
Introduction to the lithostratigraphy of the Moncton Basin
"Geo-tourism" - Hopewell Rocks
Albert Formation lacustrine oil shales and Albertite solid hydrocarbons
Core viewing: Shell Albert Mines #4, Gulf Minerals Millstream #42, Columbia G56

(KEIGHLEY)

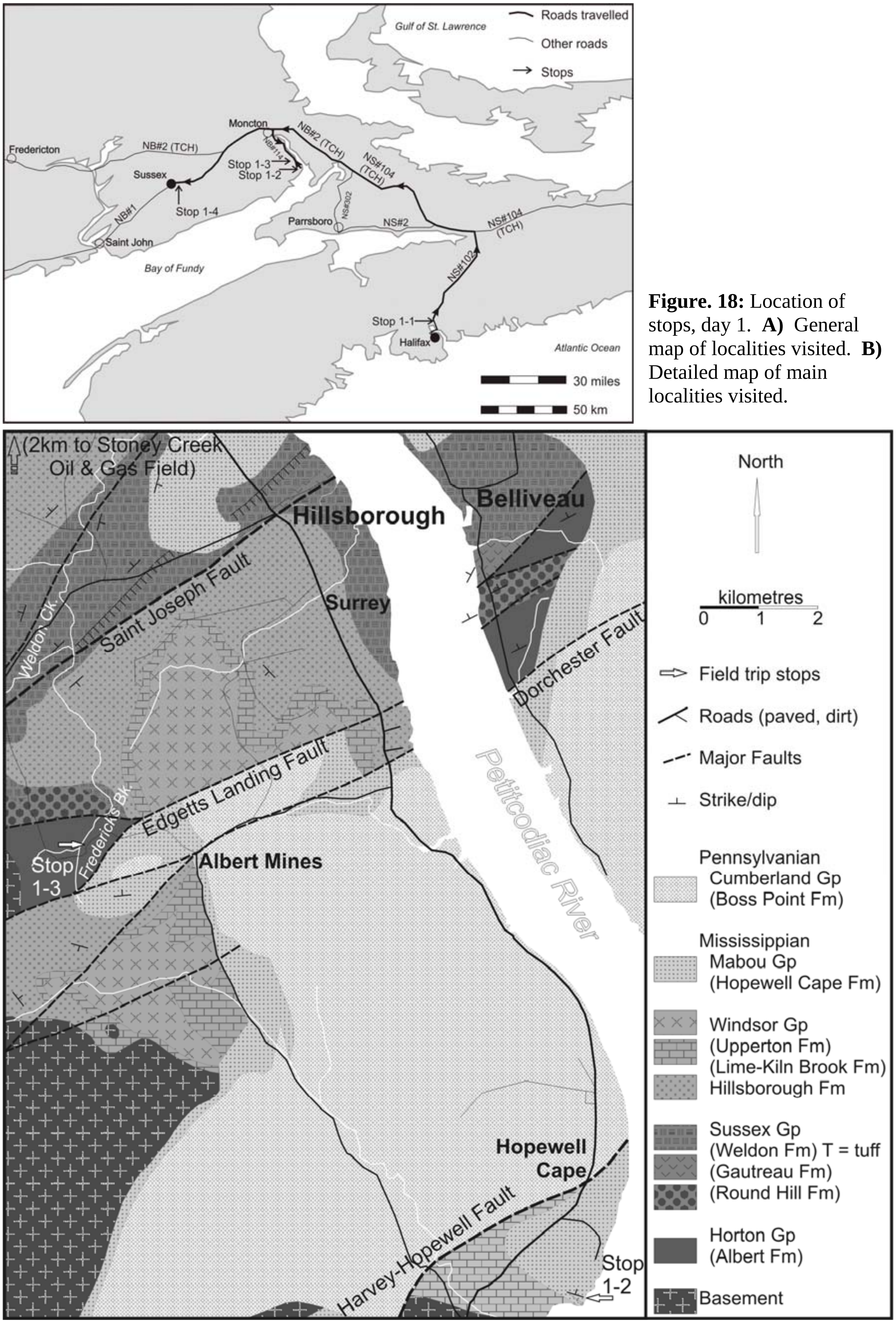


Figure. 18: Location of stops, day 1. **A)** General map of localities visited. **B)** Detailed map of main localities visited.

Convene

Location: Dalhousie University, 8.30 am

Depart 08.45am, prompt

Purpose:
Welcome. Loading of luggage onto vans. Roll call!
Please arrive in plenty of time. We have a tight schedule to meet if we are going to have time for some "geo-tourism" on the way (Figure 18).

En Route
Departing out of the Kings College gate of Dalhousie University, turn left onto Coburg Road, continue to the traffic lights and turn right onto Oxford Street. After ~0.3km turn left at first set of lights onto Jubilee Road then, after two side streets turn right onto the divided highway of Connaught Avenue. Continue along Connaught for ~2.7km to the lights where the road merges with Windsor St. Continue northward along Windsor for ~0.3km, getting in the 'straight ahead' lane signed for the McKay toll bridge. A major 5-way intersection is at the end of Windsor. Veering slightly right through the lights, follow the highway to the bridge, staying in the right hand lane. Cross the bridge and pay the toll ($0.75). Continue straight along Highway 111 then (after ~3.5 km) take the Exit 4 ramp. Get in the left lane and follow the ramp round onto Highway 118 northbound. Drive ~12 km then take exit 14. At the intersection turn right then immediately left onto Boy Scout Camp Rd. Park at the road's end.

Stop 1-1. Modern analogues, and lacustrine sequence-stratigraphic concepts

Location: Lake Thomas/Miller Lake (15 mins)

Depart 9.30

Purpose of stop:
View of two modern 'open' lakes (modern analogues).
Identification of modern depositional environments.
Introduction to sequence-stratigraphic concepts in lacustrine basins

Comments:
This stop provides an opportunity to view modern-day lacustrine basins: Lake Thomas and Miller Lake. Unfortunately, time restrictions do not permit us to view close-up any of the current depositional environments in the basin.

Both lakes currently have an outflow: they are 'open' lakes that fill all available space in their respective small catchment basins. What might happen in years of drought? What would happen if the weir at the outlet of Miller Lake was raised? What would happen if, instead of a bridge for Highway 102 across Lake Thomas, a causeway had been built?

These concepts of open and closed lakes introduce an important control on the architecture of sedimentary strata in such basins, and to sequence stratigraphy in nonmarine basins. Open lakes provide a stable base level and will usually remain open as long as the climate does not become drier. Closed lakes can undergo relative base level rise if the climate gets wetter, base level fall if the climate gets drier (see introduction).

En Route
Drive back up Boy Scout Camp Rd then right at the junction. Follow the road, sharp right after the underpass then curving back left, to the lights. Cross route 2 onto the on-ramp of Highway 102 northbound. Highway 102 merges with 118 after ~0.75 km. Continue on 102 for ~40 minutes (~73 km). After passing the town of Truro, get in the left lane (Highway 102 ends) signed to New Brunswick and Highway 104 (Trans-Canada) westbound. Stay on Highway 104 for ~ 1 hour (paying a toll ~$4.00 near half-way), until the border (total ~108 km), where the road becomes NB Highway #2.

After ~35 minutes (~60 km), take Exit 454 (Moncton, Mapleton Road). Turn left onto Mapleton. After 1 km, turn right onto Wheeler Boulevard (NB Highway 15). After ~4.5 km get into the middle lane. At the circular, take the 2nd exit (signed to Riverview) and cross the causeway and overpass. Immediately after the overpass, take the right off-ramp. At the lights, turn right onto Coverdale Road (NB Route 114) and head east. Keep straight on this road (past the left turn onto the 'narrow bridge'), which then becomes the Hillsborough Road (NB Route 114). Continue for ~35 minutes (40 km), through Hillsborough, to Hopewell Cape. Turn left and follow the road round to the car-park.

Stop 1-2. Introduction to the lithostratigraphy of the Moncton Basin. Geotourism. Lunch.

Location: Hopewell Rocks (75 mins)

Depart 13.45

Figure 19: Two views of the flower pots at Hopewell Cape **A)** looking south, **B)** looking north.

Purpose of stop:
Introduction to the lithostratigraphy of the Moncton Basin
Geotourism: "Flower-pots" at Hopewell Cape. Hopewell Conglomerate and interfingering Windsor Group limestone.
Lunch (packed lunch provided) and rest-stop.

Comments:
The famous "flower pots" at Hopewell Cape (Figure 19) are a result of weathering and erosion of the fractured red conglomerate outcrop (Hopewell Cape Formation"). The separation of the individual sea stacks is initially by weathering concentrated along fractures running vertically through the conglomerate. The thin basal necks of the stacks result from undercutting by tidal and wave action.

We shall take a path down to the south end of the cliffs and then walk upsection through the succession, encountering the limestone of the Windsor Group that interfingers with the conglomerate. The conglomerate is sourced from the Caledonian Mountains to the west. Locally, cross-stratified sandstone, and imbrication in the sandier conglomerate, suggests that the succession formed in a fluvial-dominated alluvial fan, or proximal braidplain setting. The Windsor "sea" is generally assumed to have transgressed from the east.

WARNING: The rocks are often wet and algae-covered, thus very slippery, particularly when boots are covered in thick mud. The tide, although incoming, should not be a hazard if we are on schedule (low tide 13:48UT = 10:48ADT) - access possible for 3 hrs after low tide . However, please do not fall behind, since you may become cut-off by one of the headlands. At the end of the "walk along the ocean floor" (as NB tourism likes to promote the site!), you have a significant climb up over 100 steps.

En route
After exiting the car park, turn right (northbound) onto route # 114. After ~9km, turn left onto the side road to Albert Mines. After ~4km, take the right turning, and continue ~2km to the next stop.

Stop 1-3. Oil shale, stages of deformation, solid hydrocarbons, fish fossils

Location: Albert Mines oil shale test pit

Depart 14.45

Purpose of stop:
Lacustrine strata of the Albert Fm. (distal setting).
Solid hydrocarbons (Albertite)
Stages of deformation: identifying soft-sediment deformation from tectonic features
Fossil hunting (Palaeoniscoid fish)

Comments:
After a stop to view the abandoned adit mine for the solid hydrocarbon vein of Albertite (Figure 20A), we shall examine the now quite overgrown oil-shale test pit, excavated to determine the feasibility of mining the shale for its petroleum. The main test pit itself (Figure 20B) has areas of variable grade oil shale. Three lithotypes are actually identified, that grade vertically (and likely laterally) into one another (Macauley and Ball, 1982).
(1) A dolomitic marlstone forms mostly thicker, massive, light greyish brown beds. Dolomite dominates with variable (0% to 40%) sand, silt, clay (illite), siderite, and kerogen also present. The kerogen may be finely disseminated or occur as dark brown flat to slightly wavy laminae.
(2) A greyish brown to brown (papery) clay marlstone forms mostly laminated beds, occasionally forming papery shale in clay-rich intervals. Clay (detrital and authigenic montmorillonite and illite) is dominant, with variable amounts of dolomite and kerogen, and minor zeolite (analcime) that likely formed in Na-rich alkaline waters from alteration of clay minerals or volcanic ash.
(3) The laminated marlstone contains the highest kerogen content. The beds have a marked varve-like appearance due to the alternation of organic rich layers with dolomite and calcite laminae.

Hydrogen/oxygen indices identify the kerogen to be a low maturity type 1, with unusually low oxygen content. This composition is typical of a primarily lipid rich, algal source, which is the main organic matter produced in lacustrine basins. Although extensive preservation of this organic matter suggests anoxic conditions below the sediment-water interface, the presence of fish fossils (the palaeoniscoid fish *Elonichthys* has been found in the clay marlstone on the west side of the pit) indicates at least the upper part of the water column was oxic. The varve-like character of the laminated marlstone suggests seasonal die-off of algal blooms.

The oil shale has long been recognised as being highly deformed (Wright, 1922). Park (in press, Figure 20C) has identified three phases of deformation in the rocks at the pit that include a series of soft-sediment slumps. Particularly on the roadside near a secondary test pit, lamination in the shale often shows interbedded mm-scale recumbent isoclinal folding. Load structures, pseudonodules, and rip-up breccias may also be encountered.

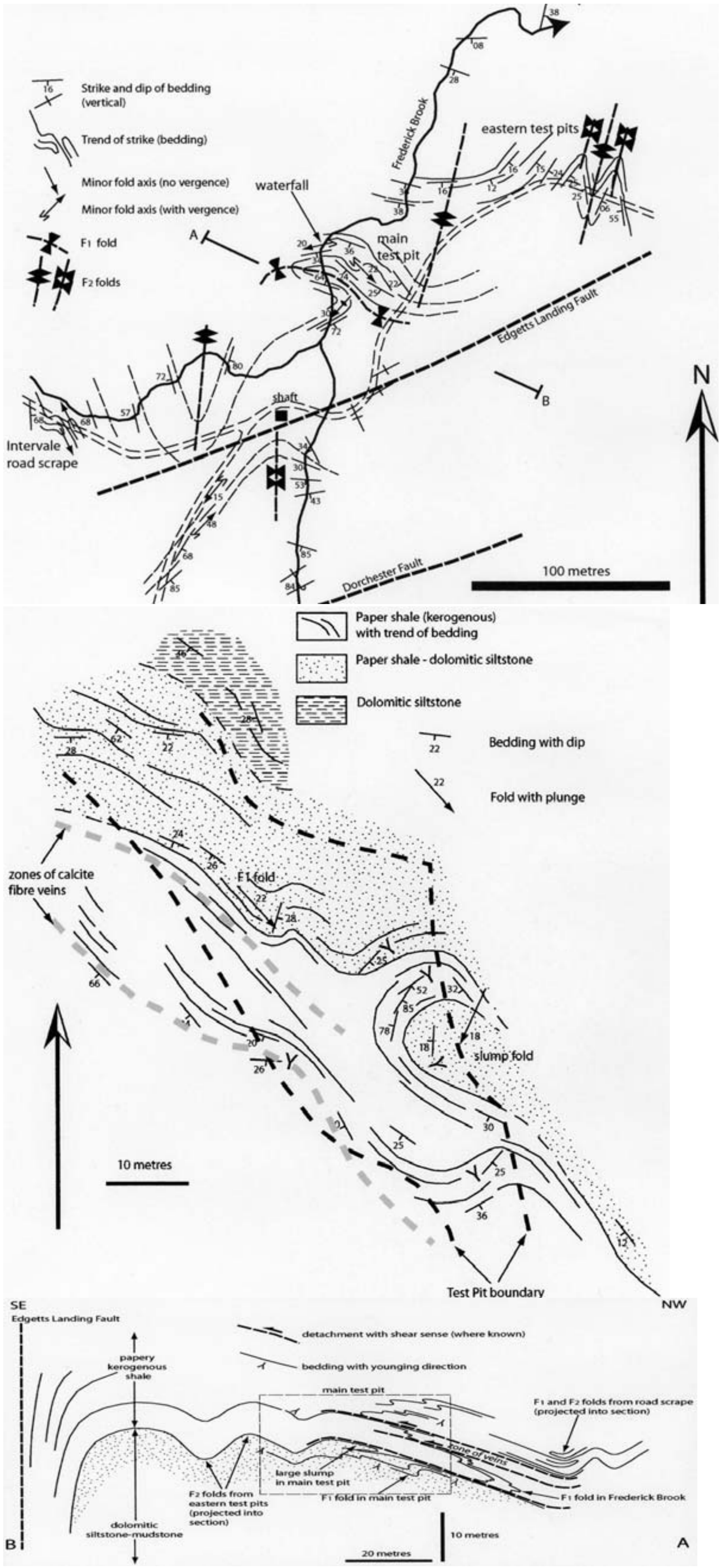


Figure 20: Maps and sections of Albert Mines (all from Park, in press). **A)** Map of the general area **B)** Map of the oil shale test pit, **C)** Interpretive cross section of the pit area (A-B in Figure 20A), viewed from the northeast.

En Route
Retrace the route back to Route 114 and north to Moncton (passing the "narrow bridge" on your right), then right turn onto the causeway. After the causeway, take the 2nd exit at the circular, onto Wheeler (NB Highway 15). After 2.5 km take Exit 3, at the intersection turning left onto westbound Berry Mills Road (Route 128). Continue ~5km to Junction 446 of NB Highway 2 (TCH), turning onto the westbound lane to Saint John and Sussex. Continue on Highway 2 for ~12 minutes (~25 km) then follow signs for Highway 1 to Saint John and Sussex (left lanes). Continue on Highway 1 for ~25 minutes (~44 km) to exit 198 signed Sussex Corner. Turn left at end of off-ramp, then left at the T-junction toward Picadilly. Follow this road for 3 km. The NB DNR field office is on your left.

Stop 1-4. Core viewing

Location: New Brunswick Dept. of Natural Resources, Sussex Field office (1 hr 45 minutes)

Depart 18.00

Purpose of stop:
View of Albert Formation core from the Albert Mines, Monteagle, and Sussex areas

Comments:
Selective intervals of core from the Maritimes Basin have been laid out for viewing. They include core boxes taken from the interval between approx. 735m to 670 m in the Shell Albert Mines # 4 well (Figure 21), and the interval between approx 400 m to 275 m in the Columbia Monteagle L18 well (Figure 22). It is also hoped to have some core from the McCully field available.

Shell Albert Mines #4 was an oil-shale evaluation drillhole that was cored from the surface to a total depth of 762.3m. The Dawson Settlement Member (641.5m, or 679.2m, to TD, depending on the pick) has only been partly penetrated by this well, but has been logged in detail where cored (e.g. Figure 21). Keighley (2000) described, between ~730.5 and ~679.0m, a major sandstone-dominated interval, enclosed by thick, dark grey, variably calcareous/dolomitic, distinct to diffusely laminated shale, readily interpreted as lacustrine facies deposited below wave-base. Shale and mudstone also interbed within this sandstone-dominated interval. They are laminated at ~729.0m and ~687.0 to ~685.0m and similarly given a quiet-water lacustrine interpretation. In contrast, the siltstone, mudstone, and associated sandstone between 704.0 and 700.3m, and 697.9 and 690.2m are bioturbated, typically massive, and retain evidence of root casts (rhizoliths). These strata are considered indicative of subaerial exposure, such as on a vegetated delta floodplain or lake margin flats where there was incipient soil development.

Keighley (2000) described the coarse-grained clastic interval as being variably sandy, pebbly, or conglomeratic, variably calcite cemented, and locally oil stained. Well-sorted, wave rippled to small-scale swaley, medium- to very-fine-grained sandstone (e.g. 728.7 to 727.2m, 720.75 to 720.0m, 717.5 to 715.4m) and any intervening soft-sediment deformed (e.g., dewatering pipes between 720.0 and 719.0m) sandstone are the confidently identified lacustrine facies. The ripples and swales indicate deposition above storm wave-base; the dewatering pipes indicating rapid deposition and subsequent sediment liquefaction. The interpretation of other sandstones is less certain. Sandstone from 693.1 to 691.7m, and 689.25 to 687.55m are medium-grained, distinctly well-sorted, and cross-stratified. Particularly in the case of the lower sandstone, its occurrence within a rooted interval might suggest deposition within a small fluvial channel. Pebbly sandstone and conglomeratic beds are more diffusely stratified by pebbly and, or carbonaceous lags, or graded strata. Beds may be between 0.5 and 2.0m thick, and fine upward. Some (e.g. 727.0 to 725.0m) may grade upward eventually into the lacustrine sandstone facies and, because of the lack of any subaerial indicators interbedded with the conglomerates, they are herein speculated to be the subaqueous part of a fan delta or Gilbert Delta. Example core intervals are shown in Figure 23. Elsewhere, similar associations have led St. Peter

(1992, 1993) also to suggest that at least some of the Albert Formation conglomerates are lacustrine, but diagnostic evidence is lacking.

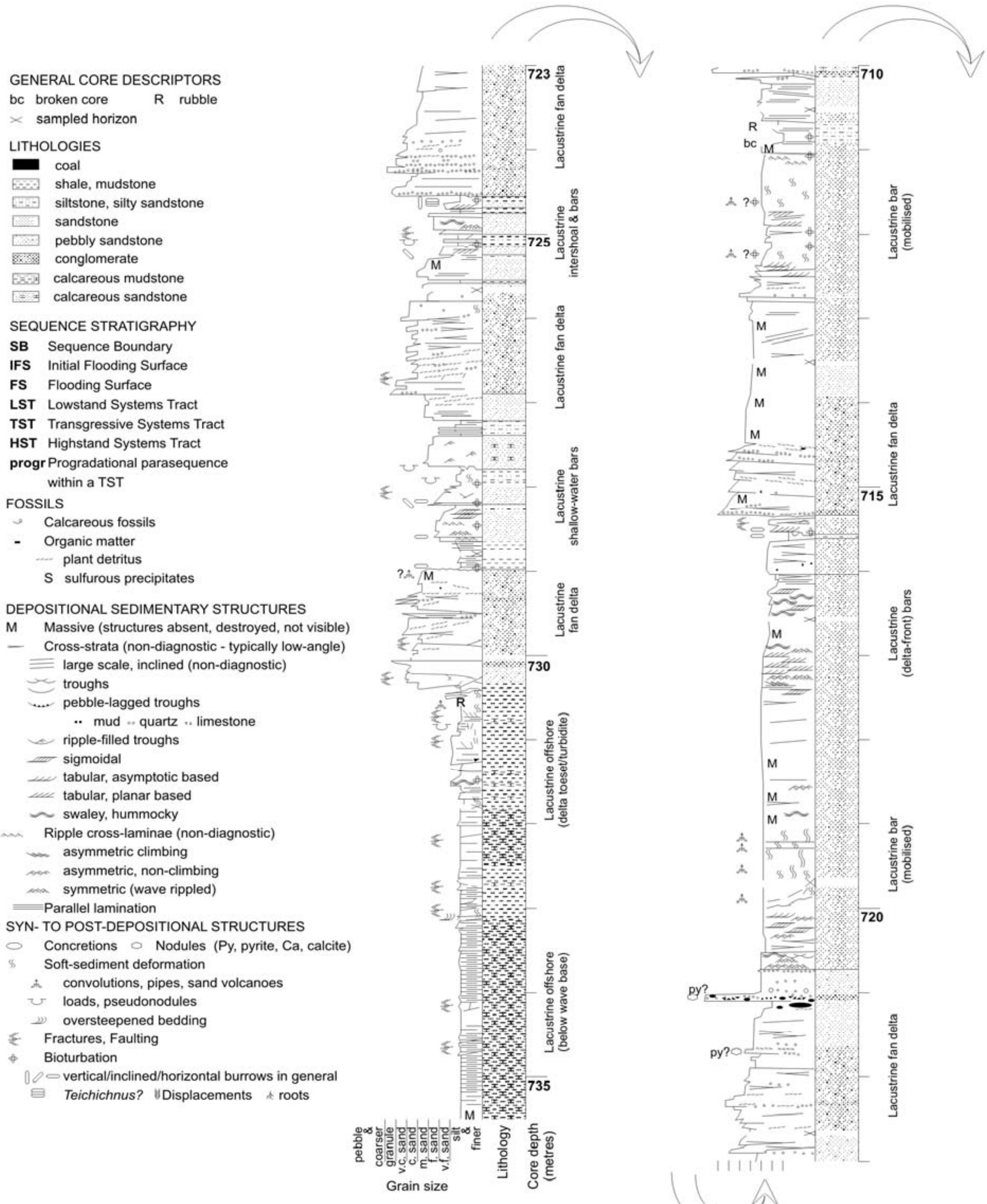


Figure 21: Part of the cored section of the Shell Albert Mines #4 well (from Keighley, 2000)

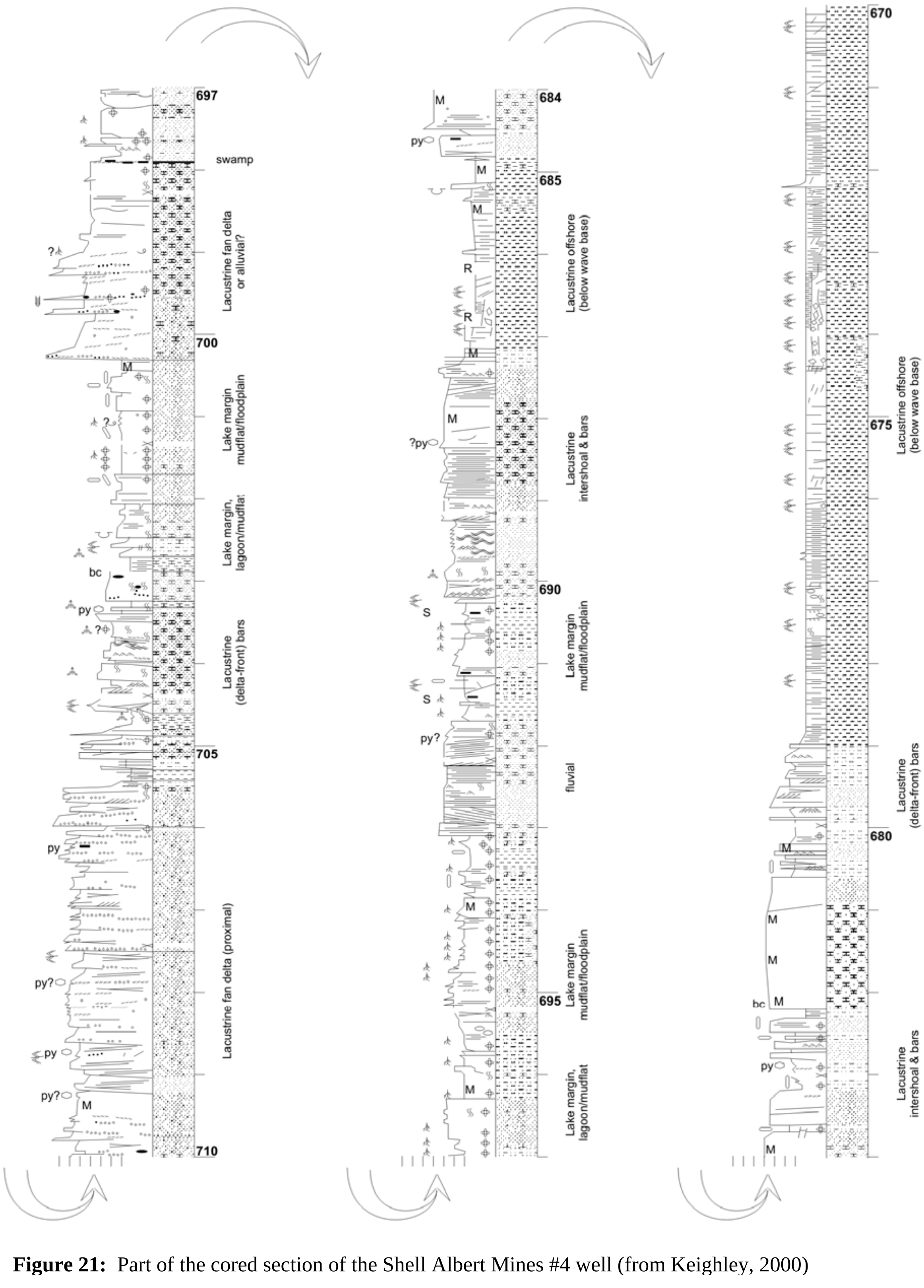


Figure 21: Part of the cored section of the Shell Albert Mines #4 well (from Keighley, 2000)

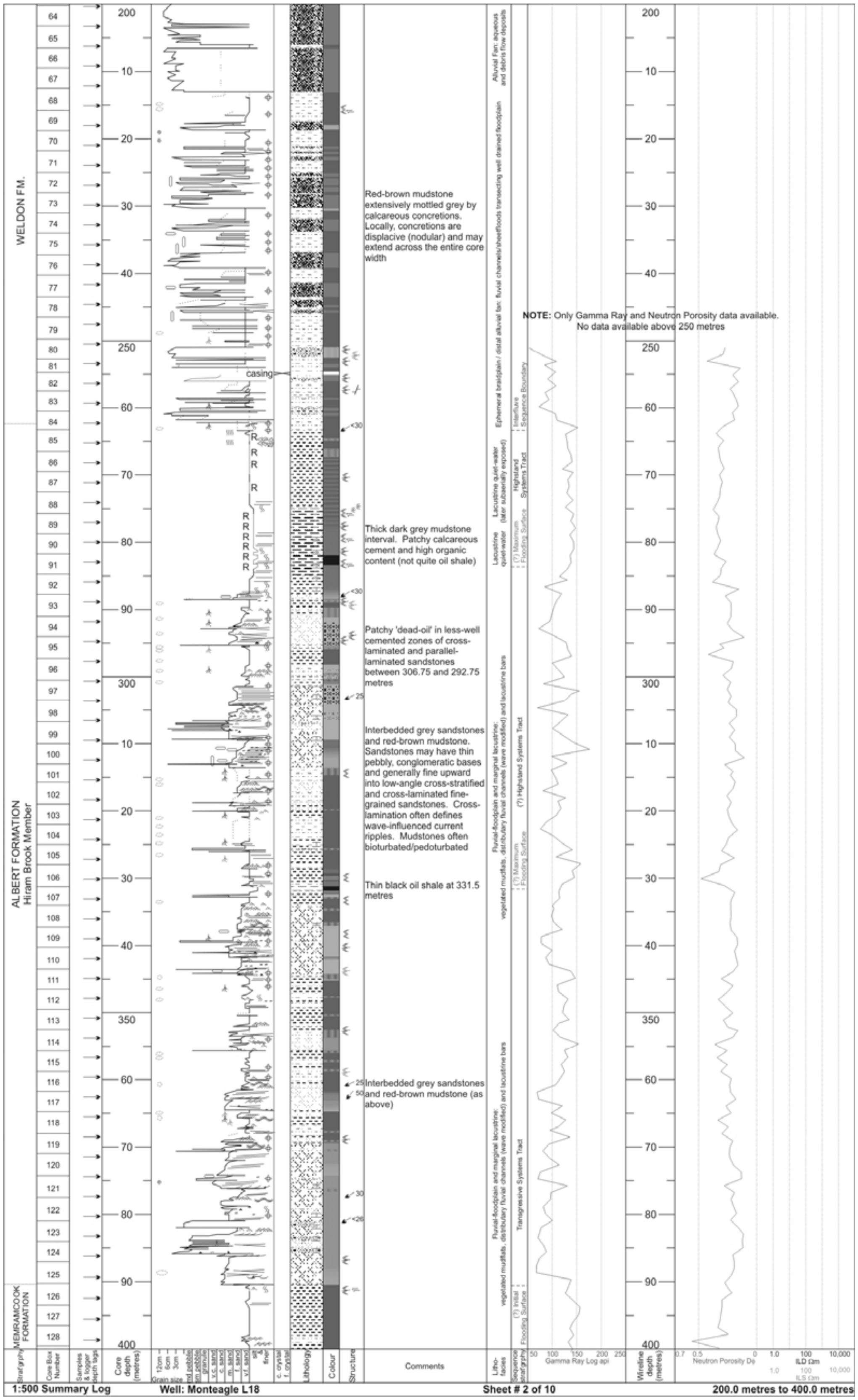


Figure 22: Part of the cored section from the Colombia Monteagle L18 well (from Keighley, 2002). Note, this log has been reduced from its original 1:500 scale.

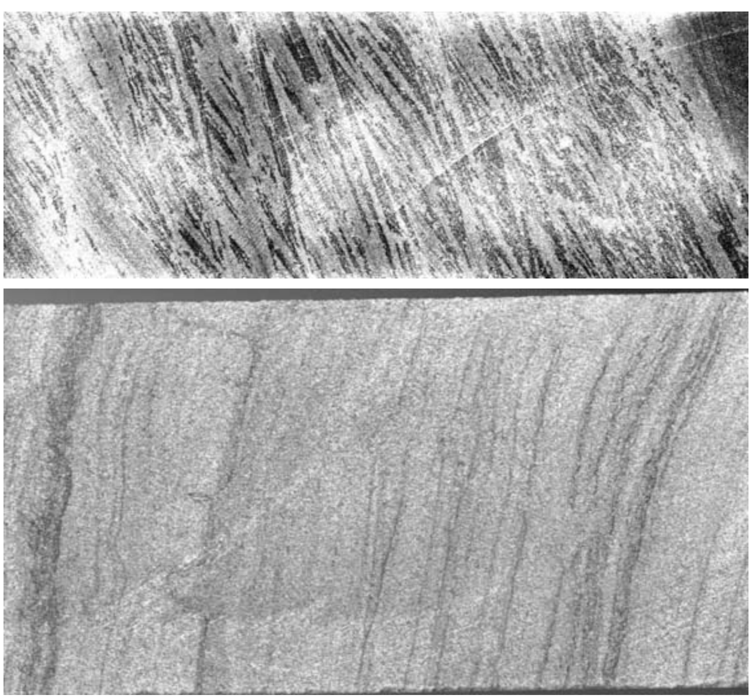


Figure 23: Core photographs. Top to the left. All core is 6cm wide. From top: Albert Mines #4, 693 m: cross-stratified sandstone. Albert Mines #4, 696.5 m: rooted, pedo/bioturbated argillaceous sandstone. Albert Mines #4, 719.25 m: dewatered, hydrocarbon stained, calcareous sandstone. Albert Mines #4, 724 m: pebbly sandstone with abundant organic fragments. Monteagle L18, 312 m: bioturbated and soft-sediment dewatered, apparently paired lamination. Monteagle L18, 299 m: dead-oil stained, swollen lens-like sets of ripple cross-lamination, suggestive of wave influence (cf. de Raaf et al., 1977).

If there is agreement on the above descriptions and facies interpretations, how would you apply a sequence stratigraphy to the succession?

Columbia Natural Resources Well L-18 was drilled northwest of Moncton, within a structural belt known as the Cocagne Graben (Figure 8). The lithostratigraphy in this belt is poorly understood, and it is less than certain (though likely) that the grey beds encountered in the borehole are Albert Formation equivalent. The cored interval of the Monteagle L-18 well is dominated by reddish grey petromictic orthoconglomerate from 1977 metres (TD) up to 548 metres (Memramcook Formation). Similar conglomerate interbeds with a varied suite of massive and laminated red-brown mudstone, fining-upward red sandstone, muddy red sandstone, massive and laminated grey mudstone above this depth. Of particular interest are the fining-upward grey sandstone, grey petromictic conglomerate and thin organic-rich mudstone and oil shale present between 390 to 265 metres (Albert Formation). Specifically, there is a thin oil shale at 331.5 metres and an organic-rich mudstone at 283 metres. What is interpreted to be dead oil is present in some of the interbedded sandstone between 306.75 and 292.75 metres (Figure 22), indicating that the oil shale and, or, organic-rich mudstone has previously been buried at least to the depth of the oil window.

Abundant grey strata are encountered that contain wave ripples (e.g. 377 metres and above) which can be considered indicative of lacustrine deposition. Also, view the apparently paired lamination at 312m (Figure 22). Could this be an indication of temporary tidal influence?

There will be an opportunity, should participants wish, to stay and view the core beyond the scheduled 'end' time. The hotel, restaurants, fast food outlets, pubs are about 10 minutes drive from the core shed. Note, attendees are responsible for their own evening meal.

En Route

Exiting the core store, turn right. Continue to the end of the road (~ 3.5 km), turn right then 1st left. Stay on this road (~2 km) passing the school and hospital, to its end. Turn right at the lights onto Main Street (NB route 121) and drive through 'downtown' Sussex for ~2 km. Continue straight on at the lights where Route 121 continues with a left turn. Almost immediately, turn left into hotel car-park.

Overnight. Sussex, New Brunswick

Amsterdam Inn, Tel: (506) 432 5050

Flaring gas from the McCully Gas Field, near Sussex, NB (Photo courtesy C. St. Peter)

GUIDE: Day 2

Outcrop relationship to the McCully Gas Field
Lithofacies of the Albert Formation
Scales of deformation in the Albert Formation
Interpreting a sequence stratigraphy

(KEIGHLEY)

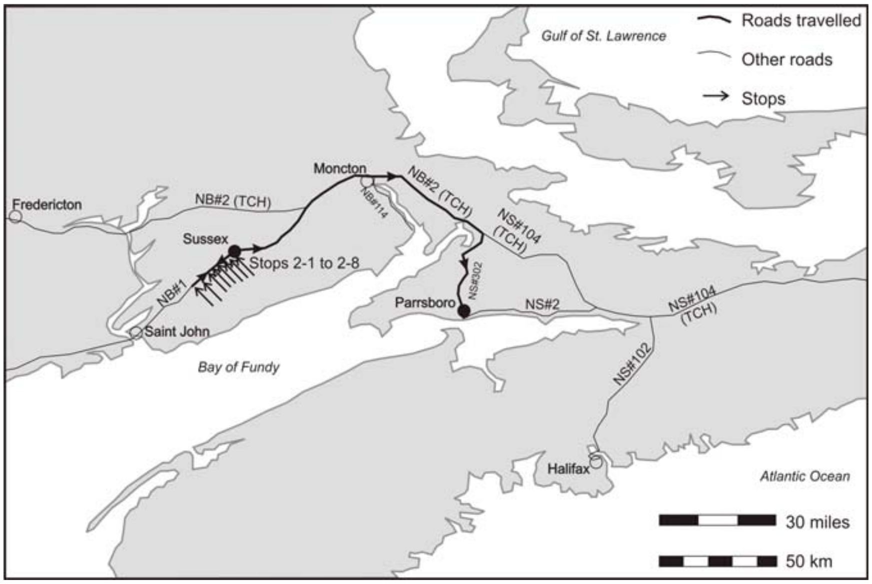
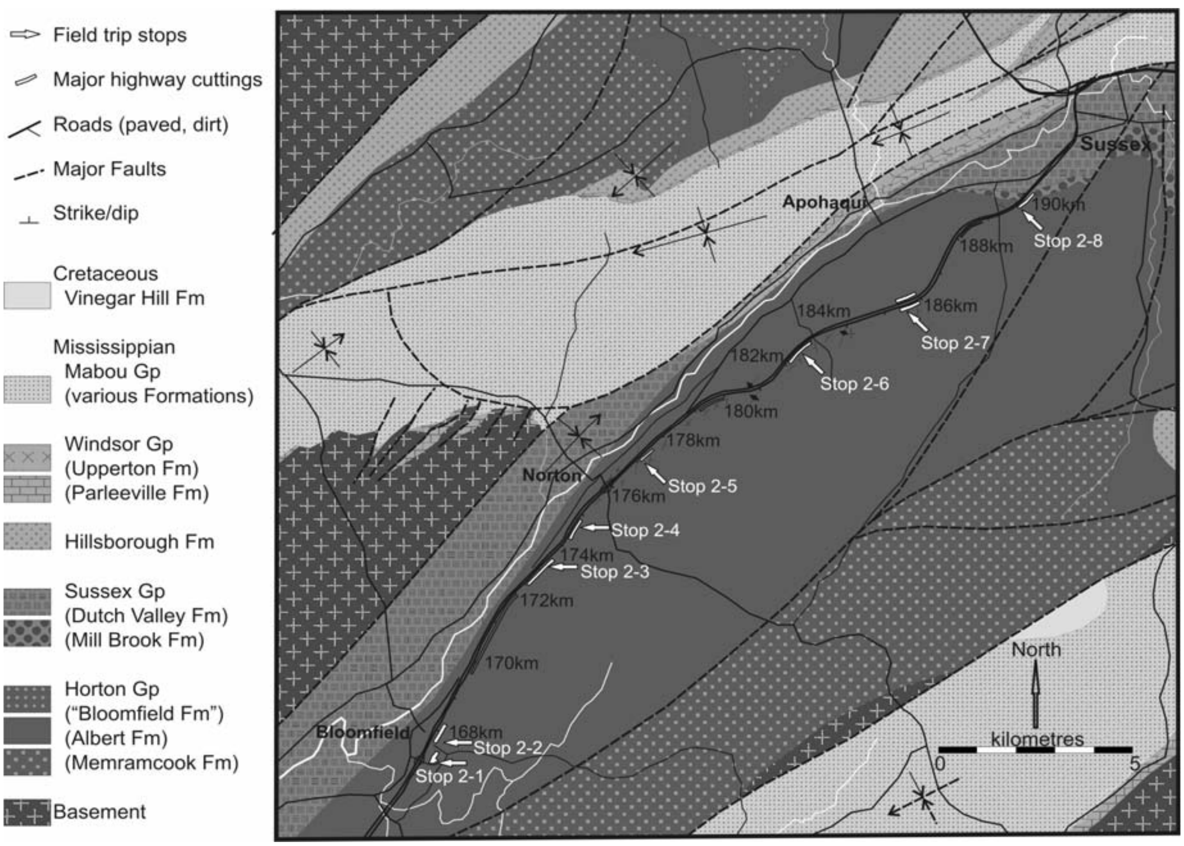


Figure 24: Location of stops, day 2. **A)** General map of localities visited. **B)** Detailed map of main localities visited (see also Figure 25).



Convene

Location: Amsterdam Inn, Sussex NB., 8.10 am

Depart 08.15, prompt

Purpose:
Check-out of hotel (continental breakfast - provided -, should be eaten by this time).

En Route
Leaving the hotel, turn right onto Main Street, then first right onto NB Route 121. Turn left onto the Highway 1 on-ramp at the intersection signed to Saint John. Continue westbound on Highway 1 for ~15 minutes (~26 km). Take exit 166 (signed Bloomfield). Turn left at end of off-ramp. After ~500 metres, turn left into the Quarry and park your vehicle.

Stop 2-1. Lithostratigraphy of the Sussex area. Outcrop:subsurface (McCully gas field) correlation, "Bloomfield Formation"

Location: Bloomfield quarry, Sussex area (50 mins)

Depart 09.30

Purpose of stop:
Comparison of outcrop lithologic variation and potential correlation with subsurface at the McCully gas field.
Sedimentology and stratigraphic position of the "Bloomfield Formation"

Comments:
This first stop will serve several purposes. First, we shall review the lithostratigraphy of the Moncton Basin, indicating lithostratigraphic differences between the Sussex area (Figure 24) and the Moncton-Hillsborough area seen 100 km to the east (day 1). Secondly, one of the geological units identified in some of the McCully wells is a red-bed unit conformably overlying the Albert Formation, and this stop is considered to have analogous outcrop. Examination of the sedimentology of this red-bed "Bloomfield Formation" unit will provide the first information to our day-long quest to produce a basic sequence-stratigraphic model for Horton Group strata in the region. Finally, the structure of the McCully gas field (east of Sussex) will be discussed, highlighting the similarities to the geological map for west of Sussex (Figure 25).

In the Moncton Basin west of Sussex, strata are considered entirely of Devonian-Mississippian age (Horton Group to Mabou Group strata - see Figure 25A). The oldest rocks are assigned to the red-bed Memramcook Formation and are overlain by a thick succession of grey Albert Formation strata to the south of the Kennebecasis River, that will be the focus of the remaining stops on this day. The grey beds are transitional upward into fine-grained red-beds that are called the "Bloomfield Formation" (although this name has yet to be formalized). These Horton Group rocks are in angular unconformable contact with overlying predominantly red-bed strata of the Sussex Group - the type area for the group being southeast of Sussex. Above another unconformity are conglomerates of the Hillsborough Formation that pass upward into carbonates and sulfates of the Windsor Group. In the subsurface, considerable thicknesses of rock salt and potash also form part of the Windsor succession in the Millstream area, Salt Springs area (mine now closed), and Penobsquis area (mine operated by Potash Corporation of Saskatchewan). Previously, the basal Windsor unit has been correlated with the Gays River Formation in Nova Scotia, although Keighley (in press) notes that the New Brunswick rocks are lithologically quite distinct, significantly thicker and, quite possibly formed part of a later marine transgression than the Gays River Formation in Nova Scotia. The Mabou Group appears in conformable succession with the underlying Windsor group, except where there has been subsurface dissolution and collapse.

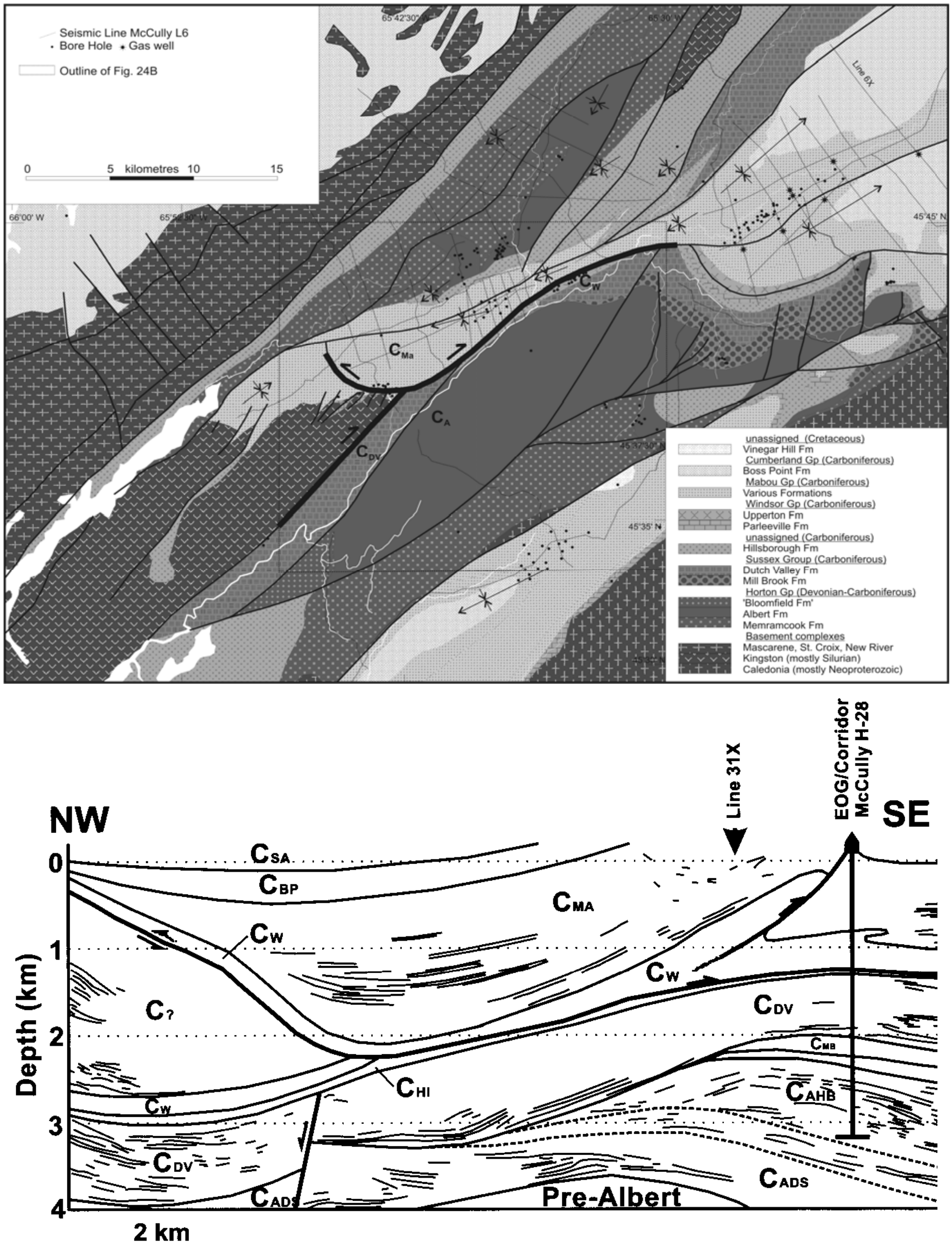


Figure 25: Geologic structure in the Sussex area. **A)** General geological map of the Sussex area. **B)** Part of the interpreted seismic line 6X, at the McCully gas field (from Wilson, 2002).

Recent work has identified two basin inversion events within the Tournaisian - one at the end of the Horton Group, the other at the end of the Sussex Group-, are identified in the Sussex area (Wilson, 2003, 2005). A later (end Mississippian) event folds Windsor Group evaporites and is associated with reverse/thrust faulting that likely includes allochthonous basement blocks (Keighley and Gemmell, 2005), while the tectonic history is concluded by Pennsylvanian sinistral displacements and reverse faulting, followed by Mesozoic normal faulting (Wilson, 2002). Further advances in the understanding of these structural events are again hampered by incomplete understanding of the lithostratigraphy, and the extent of various unconformities within the succession (e.g. St. Peter, 1993; Wilson, 2003; Park et al., 2005).

The major foundation of the above interpretation is seismic data from the McCully field (Figure 25B, Wilson 2002). Recent remapping of the area west of Sussex by Keighley (in press) has identified several contact relationships complementary with what are interpreted for the subsurface at McCully. Accordingly, if you accept such a correlation, the stops visited for the rest of the morning can be considered outcrop equivalents of the succession at McCully.

En Route

Exiting the quarry, turn right, then right onto NB Highway 1 eastbound (signed to Sussex). The remaining stops of day 2 are all roadside outcrops on the highway while heading back toward Sussex.

Stop 2-2. Sedimentology of the Albert Formation: fluvial-floodplain lithofacies

Location: Highway 1, eastbound, between 168 & 169 km markers (30 mins)

Depart 10.00

Purpose of stop:
Sedimentology of "tombstone" grey sandstone and mudstone

Comments:
The rocks exposed in this roadside outcrop are located stratigraphically immediately below the transitional interval with the overlying red-bed "Bloomfield Formation" (stop 2-1) and so represent the uppermost beds of the Albert Formation. The sandstone is coarse grained to gravelly, mostly angular, and mineralogically very immature, with a high abundance of feldspar. Cross-strata are present in lenticular sandstone beds, favouring a near-to-source fluvial interpretation for the sandstone. What is the palaeoflow direction? The sandstone is very tightly cemented, probably as a result of dissolution and authigenesis of unstable feldspar components. Cementation is particularly favoured at this locality because of its proximity to the major fault in the nearby valley, which may have acted as a major conduit for subsurface fluids.

Stop 2-3. Sedimentology of the Albert Formation: forested floodplain-marginal sandflat facies I

Location: Highway 1, eastbound, adjacent to the 173 km marker (30 mins)

Depart 10.35

Purpose of stop:
Analysis of large, horseshoe-shaped structures

Comments:
Rygel et al. (2004) recognized that the U-shaped structures commonly preserved on the bedding plane occasionally enclose fossilized trees preserved perpendicular to bedding and thus likely in-situ (Figure 26). Accordingly, the structures are considered to be scours that flowed around the impeding tree trunks, cutting down into (now weathered away) floodplain mud during a fluvial flood event.

Figure 26: Outcrop photos from stop 2-3. **A)** Horseshoe-shaped structures, **B)** Large scale straight to sinuous crested asymmetric ripples (photos courtesy of M. Gingras).

The fine to very fine grained sandstone that now moulds the scours was deposited during the waning flow of the flood. Gingras (pers. comm.), using the Hjulstrom diagram, make a rough estimate of flow velocities exceeding 60cm/second.

Rocks in the vicinity of this stop are highly deformed (witness the near vertical bedding) and their stratigraphic position within the Albert Formation is uncertain.

Stop 2-4. Sedimentology of the Albert Formation: forested floodplain-marginal sandflat facies II

Location: Highway 1, eastbound, adjacent to the 175 km marker (50 mins)

Depart 11.30

Purpose of stop:
Ancient forests and large sandstone filled cracks

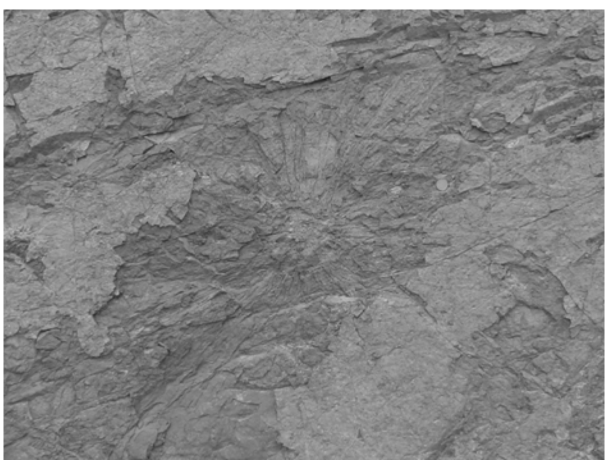


Figure 27: Outcrop photos from Stop 2-4. **A)** *Protostigmaria* fossil and **B)** large sandstone filled cracks (both photos courtesy M. Gingras)

Comments:

Falcon-Lang (2004) has recently published on the fossil forests (e.g. Figure 27) of the Norton area. This stop provides the opportunity to view the upper part of his section 4 that includes, at 14.5 metres (Figure 28), the bedding plane expressions of the mostly *Protostigmaria* (lycopsid) trees from his palaeosol 4-192 (Figure 27A). His calculations suggest a monoculture of up to 30 000 trees per hectare were present at this site - probably an upper limit above which the forest self-thinned due to

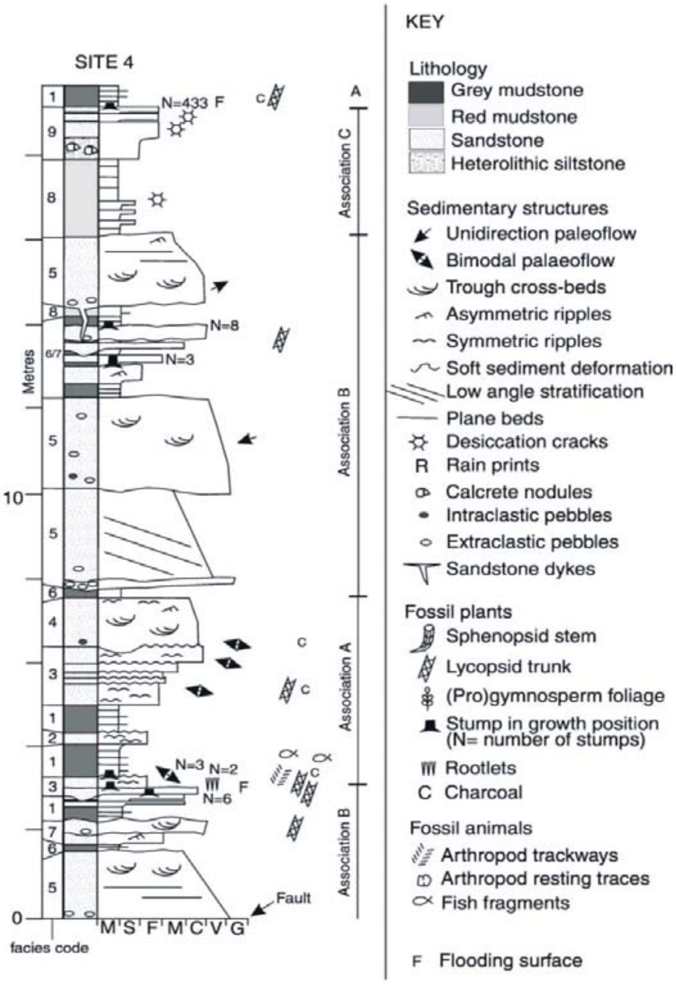
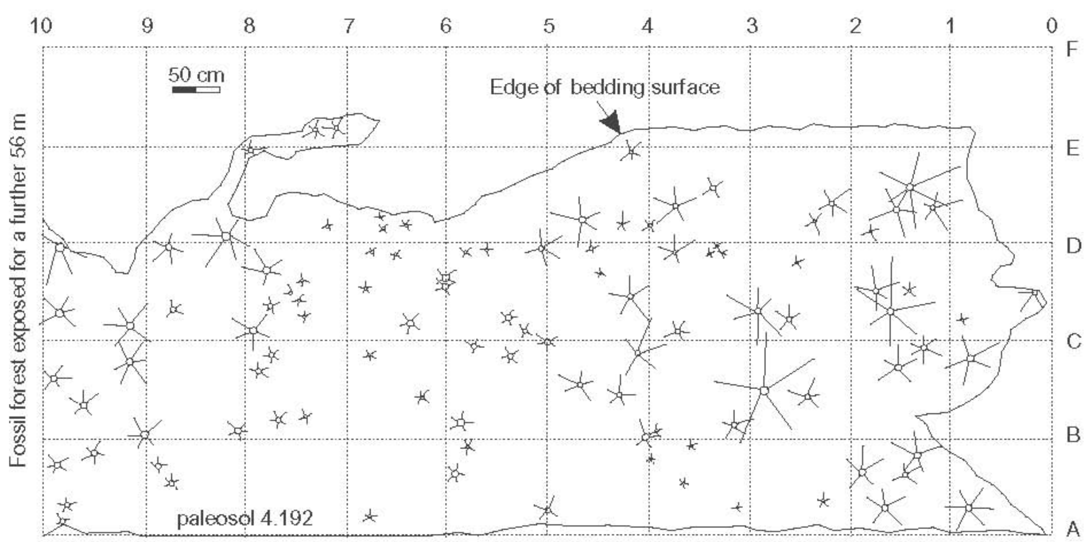


Figure 28: Detail of the Norton Fossil Forest. at the 175km outcrop. **A)** Log of Section 4. **B)** plan view of the fossil forest. Both diagrams taken from Falcon-Lang (Geol. Soc. London, 2004)



competition for nutrients. The microphyllous canopy of these forests was unlikely to have resulted in competition for light. However, forests never achieved climax because of regular flood-disturbance.

Falcon-Lang's log (Figure 28) indicates that the nearby large sandstone-filled fractures (Figure 28B) represent fissures in the sediment that were infilled downward from an overlying sandstone, though the cause is debatable. Their orientation is inconsistent with desiccation, and their scale would be remarkable for diastasis - work on such features is ongoing (e.g. Wilson, 2005).

Rest-stop/early lunch.

Location: Gas station, Norton exit of Highway 1 (45 minutes)

Depart 12.15

Stop 2-5. Sedimentology of the Albert Formation: marginal mudflat - quiet water facies, tectonic folding

Location: Highway 1, eastbound, past the 177 km marker (30 minutes)

Depart 12.45

Purpose of stop:
Albertite veins
Tectonic structure

Comments:
Outcrop (Figure 29) consists mostly of laminated grey or dark grey mudstone, grey, fine-grained, (long-crested ripple) cross-laminated, irregularly laminated, or massive sandstone (lacustrine mud/sand-flat, quiet water subaqueous). However, at both the base and top of the section are significant, cross stratified, sharp based, light grey, highly feldspathic pebbly sandstone. These units are similar to the fluvial facies identified at stop 2-2. Also present, within the laminated mudstone around 10 metres (Figure 29), are small veins of solid hydrocarbon (presumably albertite).

Rocks in the vicinity of this stop have again undergone significant deformation. According to Wilson (2004) the beds are folded into a large-scale, westward verging shallowly southward plunging anticline, with minor folds exposed in the core. They resemble folds that at later stops are interpreted as soft-sediment in origin. However, he considers them tectonic because they are not sharply bounded above and below with undeformed strata, there is a reversal of way up across the zone of minor folds, and axial plane measurements of the minor folds show point maximum distributions rather than plotting as a great-circle girdle as syn-sedimentary folds do at stop 2.7. Wilson (2004) additionally relates the deformation to the major northeast trending "Kennebecasis River Fault" located in the valley north of the highway. The overturned fold mapped at this stop trends ~020^{o} and so, if the fold is en-echelon with respect to the fault, the sense of asymmetry indicates dextral movement. Also, since Mabou Group rocks are juxtaposed on the other side of the valley, deformation must post-date them (e.g. end Mississippian transpression).

Stop 2-6. Sedimentology of the Albert Formation: marginal - offshore facies, soft-sediment deformation

Location: Highway 1, eastbound, between the 182-183 km markers (1 hour)

Depart 13.45

Purpose of stop:
Soft sediment deformation features

Comments:
At this stop we shall be walking upsection toward the southwest (Figure 30). The succession first shows an overall fining upward, then displays a broad coarsening upward. Walking quickly past the first 100 metres or so, numerous medium to coarse grained, locally pebbly, sharp based sandstone units interbed with grey mudstone. The latter are commonly crossed by large polygonal cracks, considered the result of desiccation despite the lack of red colouration in the rock. The succession is interpreted as a low lying fluvial floodplain (possibly a fluvial-dominated delta top) with periodic lacustrine innundation. The overlying thick siltstone-very fine sandstone package again shows cracks, but the regular parallel lamination suggests more prevalent quiet water lacustrine conditions.

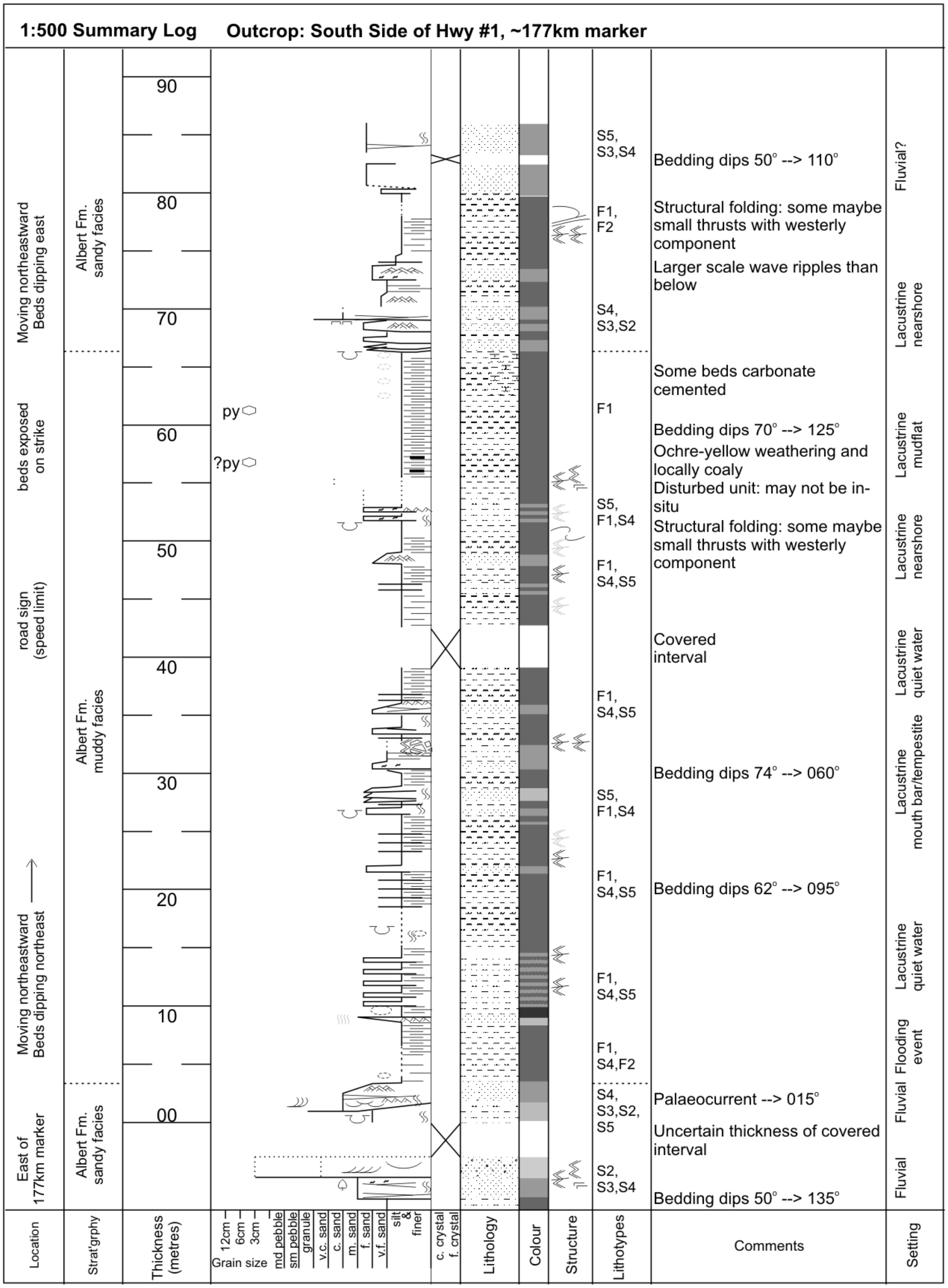


Figure 29: Rough sedimentary log of the section exposed adjacent to the Highway 1, 177km marker. Note, the log has been reduced from its original 1:500 scale.

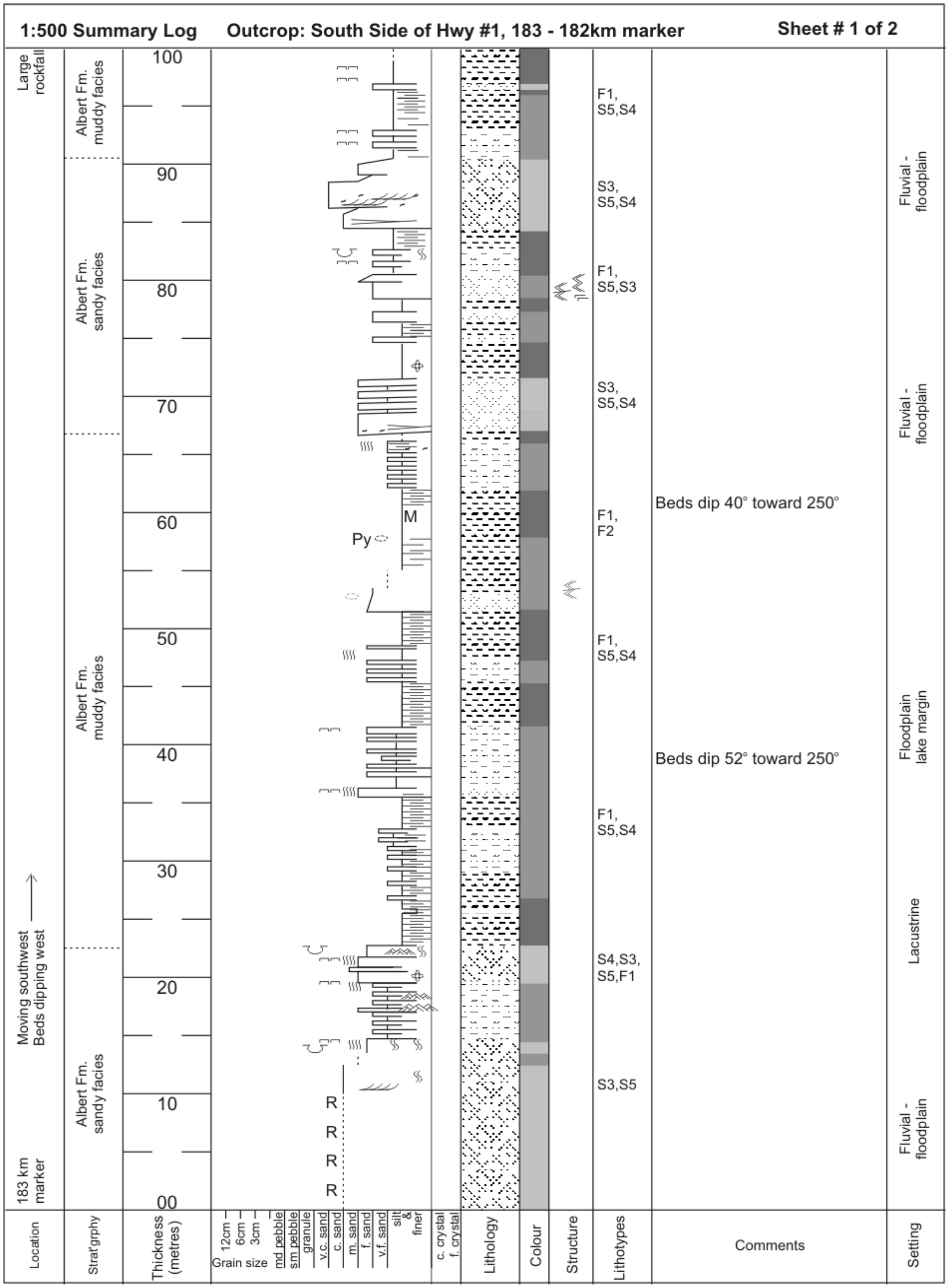


Figure 30: Rough sedimentological log for the roadside outcrop between the 183 and 182 km markers. Note, the log has been reduced from its original 1:500 scale.

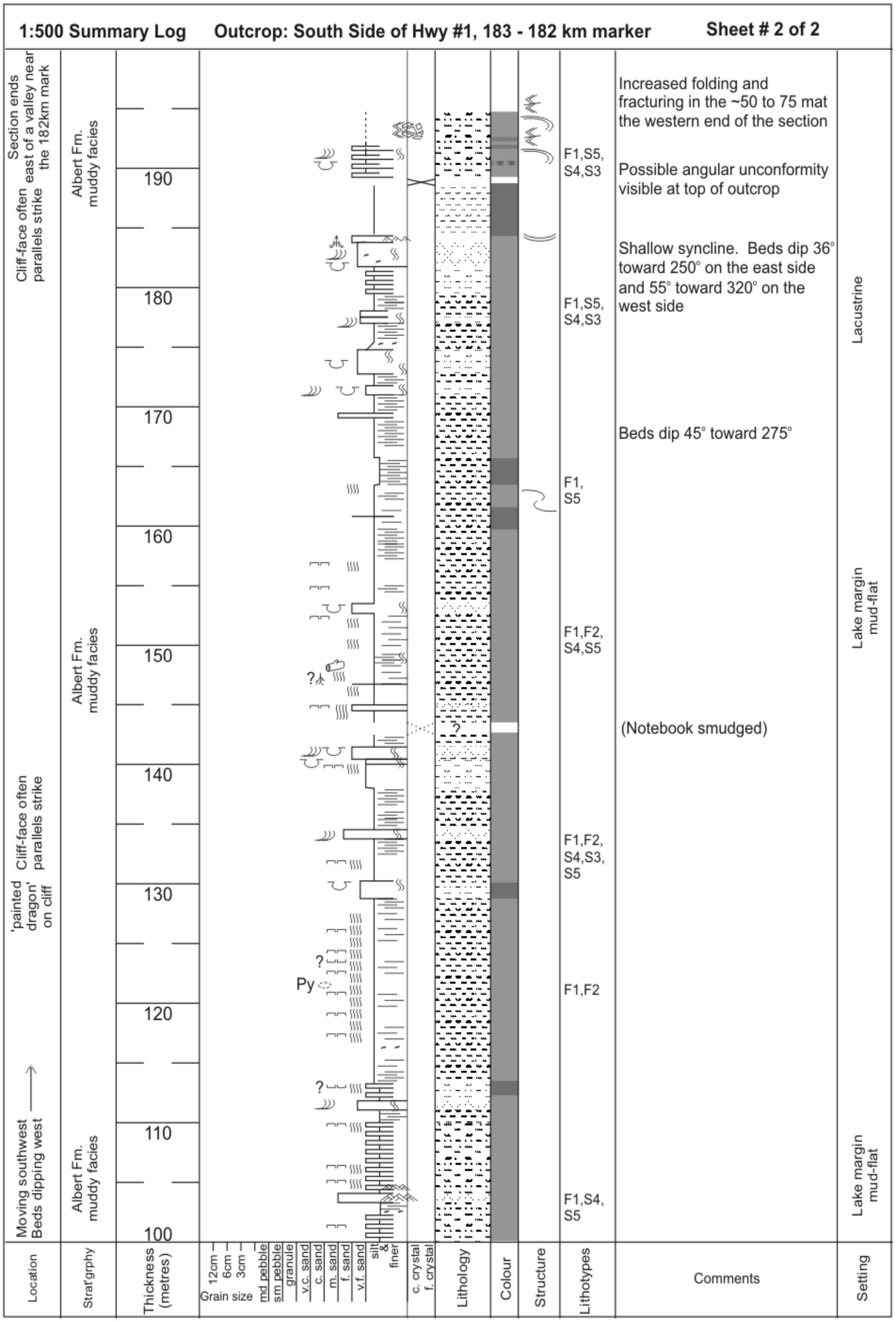


Figure 30: (continued):

Around the 170 - 180 m mark (Figure 30) are several bedding-parallel high strain zones. Refolded isoclinal folds are present (Figure 31A), with a sharp basal surface and a lumpy top surface. Overlying beds are of undisturbed laminated fine-grained sandstone, and the irregular contact indicates a sedimentary slump, rather than a tectonic origin. Close by are convolutely laminated units - were these units originally slumps that may have evolved into distal turbidites? Another bedding-parallel high strain zone that passes upward into a zone of structureless sandstone containing brecciated shale clasts and sandstone filled fissures that can be traced upward into a bed containing spectacular sand volcanoes undoubtedly formed at the subaqueous sediment surface. Similar sand volcanoes, on the top surfaces of syn-sedimentary slumps, have been described from the Namurian of Ireland by Gill and Kuenen (1957). An extensively rippled surface overlies the sand volcanoes (Figure 31B).

The next features of interest are recumbent folds with (broken) hinge-lines gently plunging toward the south. No other folds of similar orientation were encountered in this area by Wilson (2004) and their origin, presumably tectonic, remains enigmatic. Continuing southwest, a rapidly deteriorating surface contains numerous squashed-pipe like structures (Figure 31C), that were first considered to be gutter casts, then sedimentary boudins, and now to be detached and refolded remnants of thin sandstone beds (Wilson, 2004, 2005). The structures in their long direction are over 3 metres in length and in section show a cm-scale parallel lamination cut at high angle by normal-sense microfaults (Wilson, 2004). In cross-section, they are variably, circular, elliptical, aerofoil shaped, or complex, and show complex internal folds, including in one instance a sheath fold (Figure 31D, E). The enclosing rock is of fine-grained sandstone to siltstone containing a rough fissility parallel to master bedding. Thin sections show cm-scale folds with axial planes parallel to bedding (Wilson, 2004).

Stop 2-7. Sedimentology of the Albert Formation: shoreface facies, soft-sediment deformation

Location: Highway 1, eastbound and westbound, 185-186 km marker (1 hour)

Depart 14.45

Purpose of stop:
Shoaling upward succession
Soft-sediment deformation

Comments:
At this locality we shall walk eastward and upsection along the south side of the eastbound highway (Figure 32) before crossing to view outcrop on both sides of the westbound highway.

The succession on the southside (Figure 33A) shows an overall coarsening upward succession. Thick dark grey shales interbed with progressively thicker bedded buff-grey, fine-grained, sandstone containing symmetrical, long crested ripples, and locally swaley cross-strata. The sandstone around the 45 metre mark is mostly cemented by rhombic ferroan dolomite. The succession is interpreted as representing a shallowing (shoaling upward), prograding lacustrine shoreface succession. At the very northeastern end of the outcrop, a few overprinting vertical structures may represent root traces, and there is also a poorly exposed coarser grained sandstone that may be considered fluvial.

CAUTION: Take care crossing the highway. Traffic is often quite heavy and fast moving.

On the north side of the median are finely interfingering organic and dolomitic shale, with some folded bedding parallel veins of siderite. At the west end of the section, the shale and slightly reddened siltstone and sandstone is again folded. In places, detached fold hinges and aerofoil-shaped hinge fragments of sandstone are present (Figure 33B), offering an illustration of how the 'boudins' identified at the previous stop originated (Wilson, 2004; Wilson et al., 2004).

Crossing to the north side of the westbound lanes, the soft-sediment nature of the folding is apparent where refolded folds are truncated by undeformed strata (Figure 33C, Wilson, 2004).

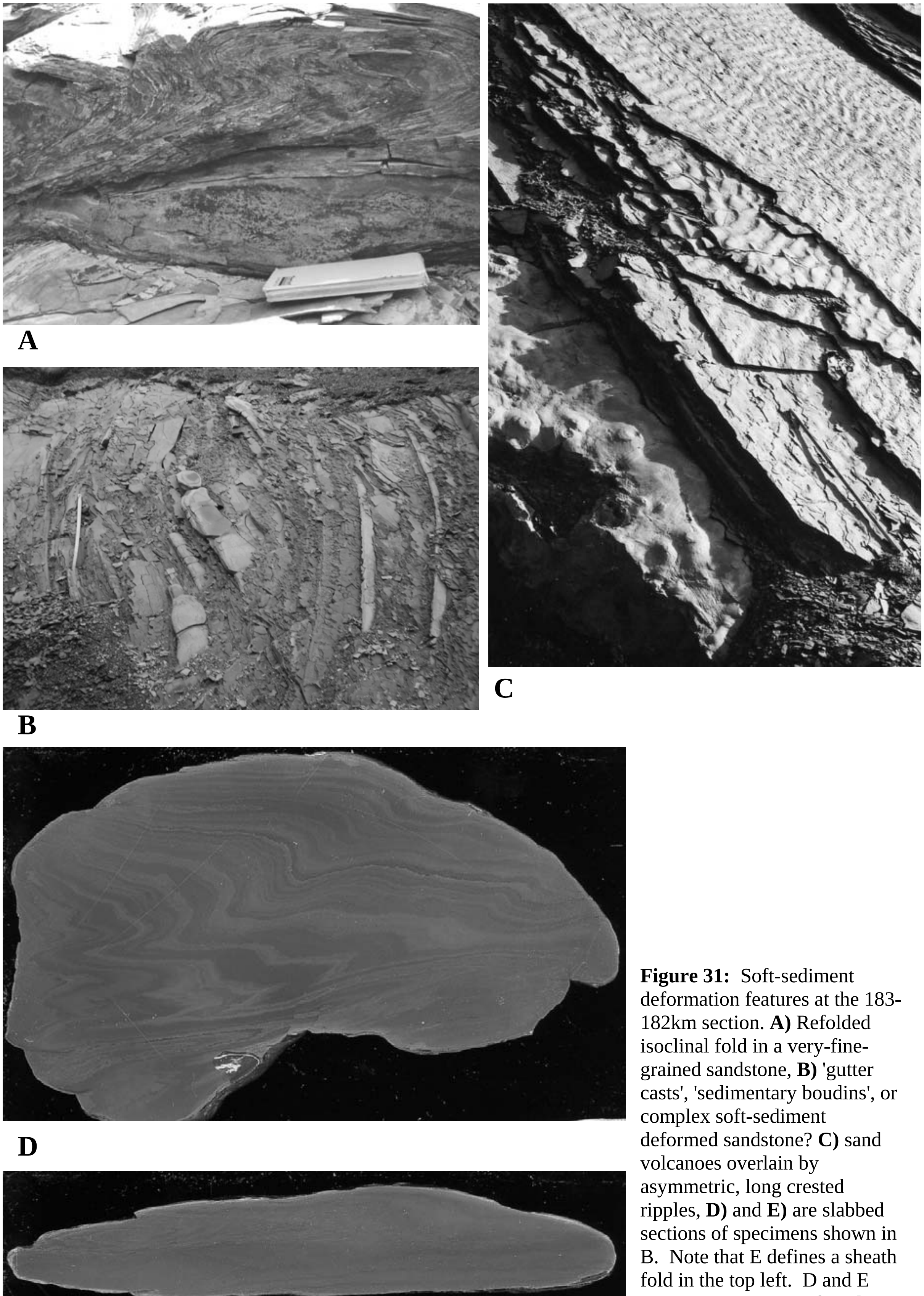


Figure 31: Soft-sediment deformation features at the 183-182km section. **A)** Refolded isoclinal fold in a very-fine-grained sandstone, **B)** 'gutter casts', 'sedimentary boudins', or complex soft-sediment deformed sandstone? **C)** sand volcanoes overlain by asymmetric, long crested ripples, **D)** and **E)** are slabbed sections of specimens shown in B. Note that E defines a sheath fold in the top left. D and E specimens courtesy of Paul Wilson

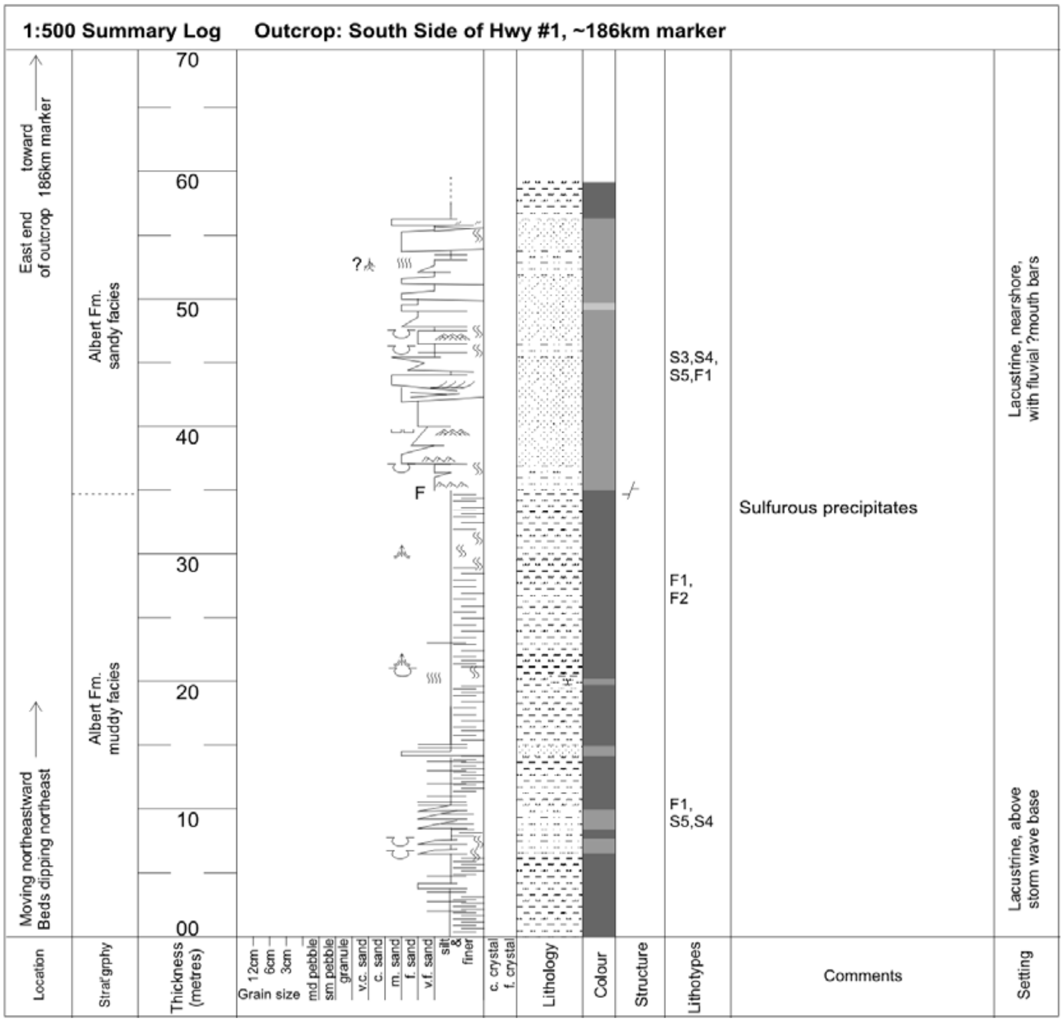


Figure 32: Rough sedimentological log for the roadside outcrop between the 185 and 186 km markers. Note, the log has been reduced from its original 1:500 scale.

Stop 2-8. Sedimentology of the Albert Formation: toward a sequence stratigraphic model

Location: Highway 1, eastbound, 190 km marker (45 minutes)

Depart 15.30

Purpose of stop:
Sussex-Horton unconformity

Comments:
This is the last stop of the day, and illustrates the contact between Albert Formation grey, locally mottled (bioturbated) purple, mudstone and very-fine grained sandstone, and tho overlying grey, polymictic pebble and cobble-grade conglomerate and pebbly sandstone of the Hazel Hill Member of the Mill Brook Formation (Sussex Group). Although apparently disconformable at this location, St Peter (2003) has been able to map the contact eastward as an angular unconformity. Slickensided calcite veins indicate that contact may have been re-activated as a fault (Wilson, 2004). Abundant pyrite is present in veins within the conglomerate.

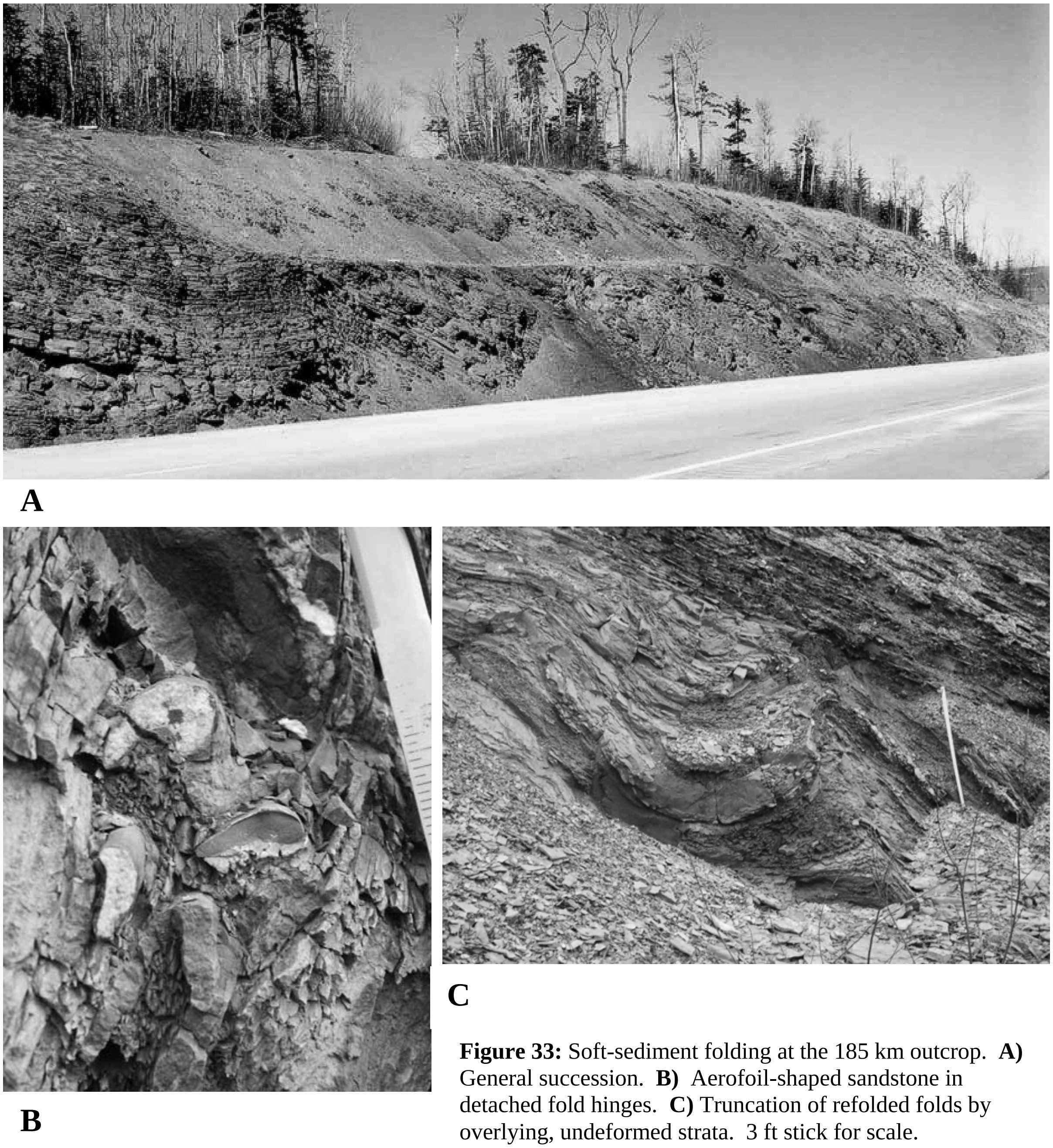


Figure 33: Soft-sediment folding at the 185 km outcrop. **A)** General succession. **B)** Aerofoil-shaped sandstone in detached fold hinges. **C)** Truncation of refolded folds by overlying, undeformed strata. 3 ft stick for scale.

Day 1 and 2 summary

At this point we shall review the various highway stops and come to some conclusions as to the facies associations that have been encountered. Obviously, with such extensive deformation, and limited vertical section in outcrop, a sequence stratigraphic model cannot be directly applied to the outcrops. However, based on what we have seen, we should be able to come to some conclusions as to what controls are most likely to have been operative.

The lack of red-beds, aeolian, and evaporitic strata precludes an underfilled sequence stratigraphic model during Albert Formation deposition. Thick successions of often slumped mudstone/shale (particularly around Albert Mines), and preservation of shoaling upward successions (e.g. stop 2-6), suggests fairly deep lakes persisted adjacent to the footwall block. Similarly, the abundance of slump folding would fit best with a significant subaqueous gradient.

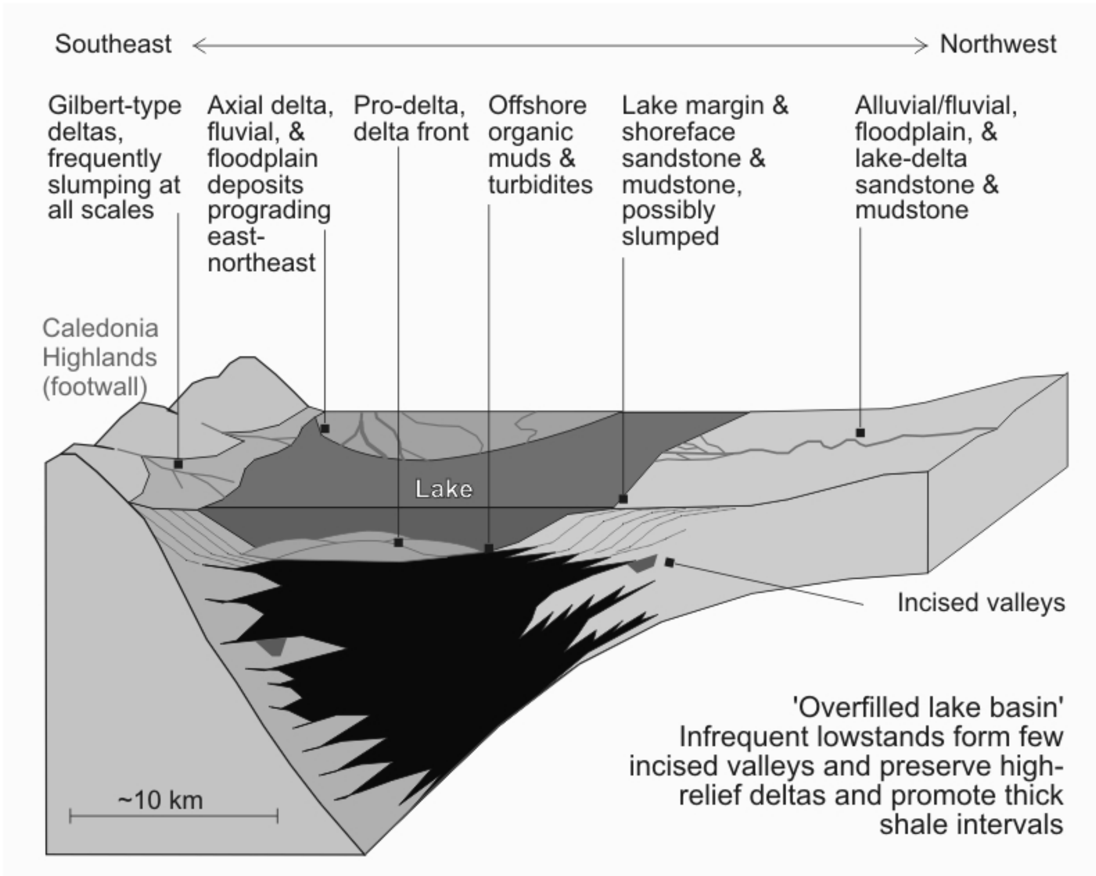


Figure 34: Suggested model for Albert Formation lakes in southern New Brunswick

The core at Albert Mines #4, however, might indicate the presence of an incised valley fill succession. Accordingly, an overfilled to balanced fill model may be perceived as the best fit (Figure 34). Most of the viewed strata likely lie above the main oil shale interval (i.e. Hiram Brook Member) which would underlie the countryside to the south, and as such represent the infilling (Lambiase Stage 3) phase of the lake. Palaeocurrents indicate mostly an easterly flow and thus axial infilling of the basin from the west (slumps likely propagated in sediments accumulating adjacent to the footwall block to the south). The "Bloomfield Formation" would represent the advancing fluvial-floodplain environment behind the axial infilling 'delta' (Lambiase stage 4).

Contrast the lithofacies seen in this first part of the field trip with what is viewed on day 3.

En Route:

Continue on Highway 1 passing Sussex, heading toward Moncton. After ~30 mins (~50 km), Highway 1 merges with Highway 2 (Trans-Canada Highway - TCH). Continue eastbound on the TCH for ~95 km, passing Moncton and Sackville. At the provincial border, the TCH becomes NS Highway 104.

In Nova Scotia, continue on the TCH for ~6 km, passing the town of Amherst. Take Exit 4, turning right at the end of the off ramp onto NS Route 2. At ~3km, turn right onto NS Route 302 toward Nappan, Maccan and Southampton (~30 minutes, ~26km) before turning right and rejoining Route 2. After ~ 25 minutes (~26 km) arrive in Parrsboro.

Overnight: Parrsboro, Nova Scotia.

Location: Sunshine Inn, Parrsboro. Tel: 902 254 3135

A spectacular view from Old Wife Point of the contact between the latest Triassic Blomidon Formation playa-lacustrine sediments and earliest Jurassic North Mountain Formation tholeiitic basalts. The Triassic-Jurassic boundary is located near the base of the light coloured bed beneath the basalts. Photo courtesy of Gela Crane (EnCana).

GUIDE: Day 3

Fundy Group stratigraphic succession
Syndepositional tectonism
Lacustrine-playa cyclic deposition
Aeolian dune field

(BROWN)

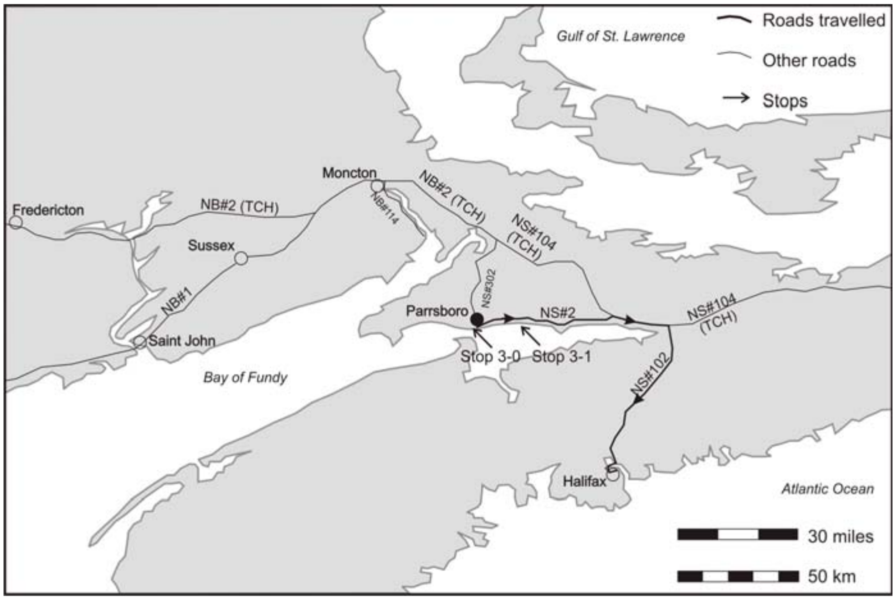
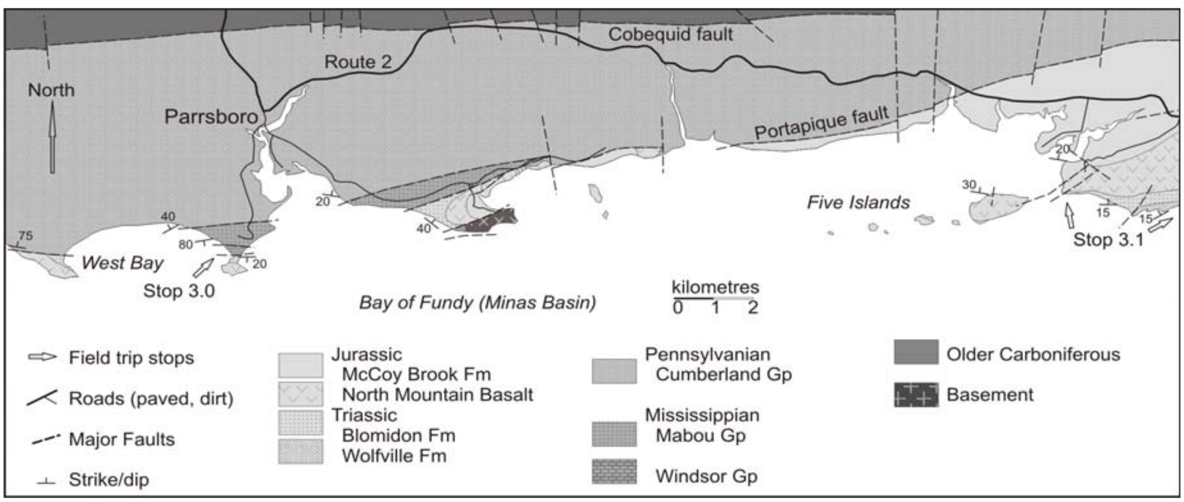


Figure 35: Location of stops, day 3. **A)** General map of localities visited. **B)**. Geological map of the northern Minas Basin area.



Convene

Location: Sunshine Inn, Parrsboro NS.

Depart 08.45, prompt

Purpose:
Check-out of hotel (Please have your continental breakfast - provided - eaten by this time).

En Route
From the Sunshine Inn, drive south ~1.5 km on Route 2 (Upper Main Street). Follow Main Street through the intersection with Route 2 (left to Halifax-Truro), past the bandstand, to its end. Veer right to head southwest along Whitehall Road. Drive 3.5 km with Partridge Island looming up ahead. Just past a sharp right turn, turn left at the sign for Ottawa House and drive down to the beach. Follow one of the better tracks to the right and park adjacent to the beach. (Figure 35, 36).

Stop 3-0. (Optional) West Bay Formation.

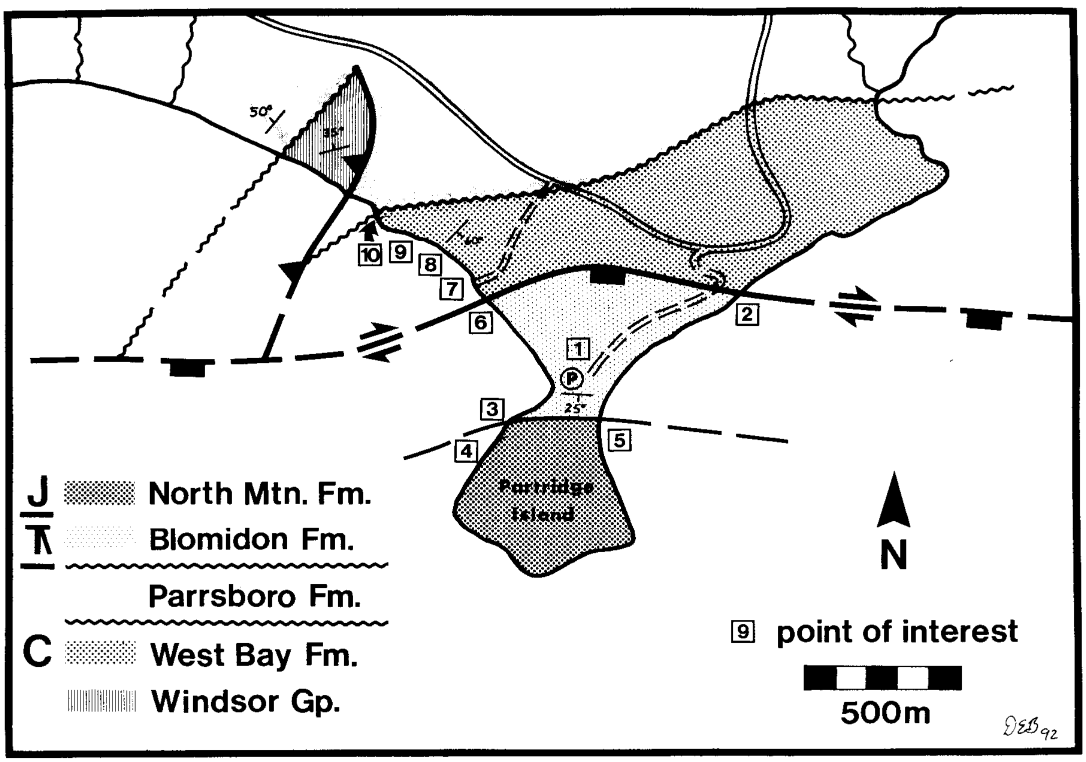


Figure 36: Simplified geological map of the Partridge Island - East Bay area

Location: East Bay, Partridge Island (45 minutes)

Depart 09.45

Purpose of stop:
Amphibian footprints / trackways (West Bay Fm.- Mississippian).
Primary sedimentary structures (West Bay Fm.)
Lacustrine facies architecture (West Bay Fm.)
Angular unconformity (Parrsboro and West Bay Fms.)

Comments:

This is an optional stop (access to our main stop at Five Islands requires the tide to ebb until ~10.30).

Partridge Island is the eastern of two headlands which bound and define West Bay, both of which are composed of Jurassic age North Mountain Formation basalts. The island is not a true island in that it is actually connected to the mainland by an impressive gravel tombola which can be inundated during the highest of spring tides (1).

Structurally, the Mesozoic formations at this stop are separated from the steeply dipping lacustrine strata of the Carboniferous (late Namurian) Parrsboro Formation by the Potapique-Parrsboro Fault (2) (D'Orsay, 1986). The fault is a dominantly east-west sinistral transverse fault with a normal component down to the south, thus preserving the younger Mesozoic rocks.

The North Mountain tholeiitic basalts conformably overlie red playa mudstones and siltstones of the Triassic Blomidon Formation, with both formations dipping gently to the south. The contact between these formations is well-exposed along the cliffs on the northwestern side of the island (3). Palynomorphs recovered from the light greenish-grey Blomidon sediments immediately beneath the basalts indicate that this thin bed most likely defines the Triassic-Jurassic boundary (Cornet, 1977). Walking south and moving upsection into the lowest flow unit, the basalt becomes vesicular in nature, and within the numerous vugs and fissures can be found several varieties of zeolites, such as chabazite, natrolite, mesolite, heulandite, apopholite and stilbite, as well as minor calcite, agate and amethyst (4, 5) (Sabina, 1964).

The Late Mississippian (Visean-Naurian) age West Bay Formation strata at this stop are famous for their well-displayed and spectacular primary sedimentary structures, and the abundant amphibian trackways preserved in shoreline sediments. These shales, siltstones and minor sandstones were deposited in depressions adjacent to, and immediately south of, the then dextral Glooscap-Cobequid strike-slip fault system. These depressions were the result of rift basins formed under a compressive regime, as opposed to those formed in the extensional Triassic-Jurassic episode. Continued motion of the fault is reflected in the compressive folding and faulting of the West Bay Formation, and the strong angular unconformity with the slightly younger fluvial sediments of the Early Pennsylvanian Parrsboro Formation (latest Visean-early Namurian).

Originally, the preserved sedimentary structures, stratigraphic sequence and trace fossil evidence within the West Bay Formation were thought to be indicative of a lacustrine setting (Carrol et al, 1972; McCabe and Schenk, 1982). Based on a detailed study of this formation and bounding strata, D'Orsay (1986) believes that the formation represents a transition from distal alluvial fan-alluvial plain to an intertidal-supratidal mudflat and estuarine sequence. Thus the West Bay Formation is interpreted to represent an emergent, arid-zone coastal alluvial fan, shoreline, and estuarine complex (likely related to that seen at Hopewell Cape on Day 1).

The Portapique-Parrsboro fault separates the Carboniferous strata from the down-dropped Triassic-Jurassic section at Partridge Island (6). Approaching the outcrops of the West Bay Formation from the east, the strata are observed to be in an almost vertical orientation, striking southwest with dips 60°-80° to the southeast (7). The resultant "flatirons" expose at least 225 metres of lacustrine sediments at this site (D'Orsay, 1986). D'Orsay (ibid), in his excellent sedimentological study, detailed the numerous facies within this formation and other associated formations. Four main deposition environments are represented in the West Bay Formation, and for simplification purposes have been informally designated by McCabe and Schenk (1982):

(1). lacustrine ("wavy bedding" facies - grey mudstones and sandstones),
(2). lacustrine margin ("red and green mudrock"facies),
(3). fluvial-deltaic ("cross-laminated sandstone" facies) and
(4). shoreline and lag deposits ("coarse-grained" facies - sandstones).

D'Orsay (ibid) interprets the lacustrine strata to be of marine origin. All four facies will be observed at this stop.

The West Bay strata are famous for the well-preseved and displayed primary sedimentary structures (8). They are best displayed in the red and green mudrocks which were deposited (and later physically altered) at the margin of a large lake. The mudstones are generally unlaminated or contain parallel horizontal or undulatory laminations. Carrol et al (1972) give an extensive, but by no means exhaustive listing of the primary sedimentary structures observed in this and the other facies. The major structures include abundant desiccation cracks, oscillation and current ripples, rain ands hail imprints, vertebrate footprints, slump structures, etc. D'Orsay (1986) has indicated that the presence of vertical and polygonally-orientated mosaics of clastic dykes and boudins may reflect the disruption of uncompacted sediments in response to earthquake activity.

Amphibian footprints are very common in the West Bay sediments (9). They occur as natural casts on the undersides of of sandstone beds, having been formed on the surface of thin, soft black mud layers which erode easily. The footprints range in size from about 1 cm to 10 cm and tail drag marks are occasionally preserved. A detailed review of footprints recovered from Carboniferous sediments throughout Nova Scotia, including those from the West Bay Formation, is given by Sarjeant and Mossman (1978).

Time permitting, the well-exposed angular unconformity between the West Bay and the red floodplain sandstones and shales of the Parrsboro Formations will be visited (10). A 2-6 metre thick sequence of coarse lithic wacke and conglomerate overlies this unconformity and marks the base of the overlying Parrsboro Formation. A variety of clasts are observed in this unit, mostly red and green sandstones and siltstones, apparently derived from the underlying West Bay beds. This unconformity is unique in that it is the only one recognised to have been formed in the Maritimes at this time (Visean-Naurian) (McCabe and Schenk, 1982; D'Orsay, 1986). The cause of faulting and folding of the West Bay sequence and the formation of the unconformity was most likely due to the areas proximity to, and repeated movement of, the mainly dextral Cobequid-Gooscap Fault and its associated splays.

En Route

Retrace route back to the bandstand/town square and follow Route 2 about 25 km east towards Truro passing through the hamlets of Moose River and Lower Five Islands and village of Five Islands. Whilst driving through the latter two communities, the earliest Jurassic-age basalt-capped Five Islands are visible to the right and from west to east are named Pinnacle, Egg, Long, Diamond and Moose Island respectively.

At the village's eastern edge, the road crosses Beaver Brook. From this short bridge begin a steep ascent up the western flank of Economy Mountain. About 800 metres up the road on the right is the entrance to the Province of Nova Scotia's Five Islands Provincial Park. To the north (left) the Cobequid Highlands, an Avalon composite terrane, should be visible across the fields. The terrane-bounding Glooscap/Cobequid/Portapique fault system ("Minas Geofracture" of Keppie, 1982) defines the southern boundary of this massif and is represented by the east-west trending valley at the base of this horst. The valley floor is underlain by the Jurassic McCoy Brook Formation. It is separated from the Mississippian Parrsboro Formation to the north by the Portapique fault. The Parrsboro Formation is manifested by a series low relief hills that rest against the foot of the Cobequid Fault scarp further north.

Turn right to enter the park and drive about 2.5 km along the paved road to the main park entrance gate and buildings, follow the signs to the picnic area and park in the designated lot near the beach. A long row of clean outdoor toilets and changing rooms are to your left. Walk across the grass down to the beach, turn left and head southeast (Figure 37).

Stop 3-1. Lithostratigraphy of the Fundy Group. Outcrop of the entire succession.

Location: Five Islands Provincial Park (4.5 hours)

Depart 3.00pm

Purpose of stop:
Superb exposures of the entire Fundy Group succession
Reflection of palaeoclimate in sediments and stratigraphy
Under-filled lacustrine basin stratigraphic architecture
Syndepositional tectonism
Fossil hunting (Semionotid fish)

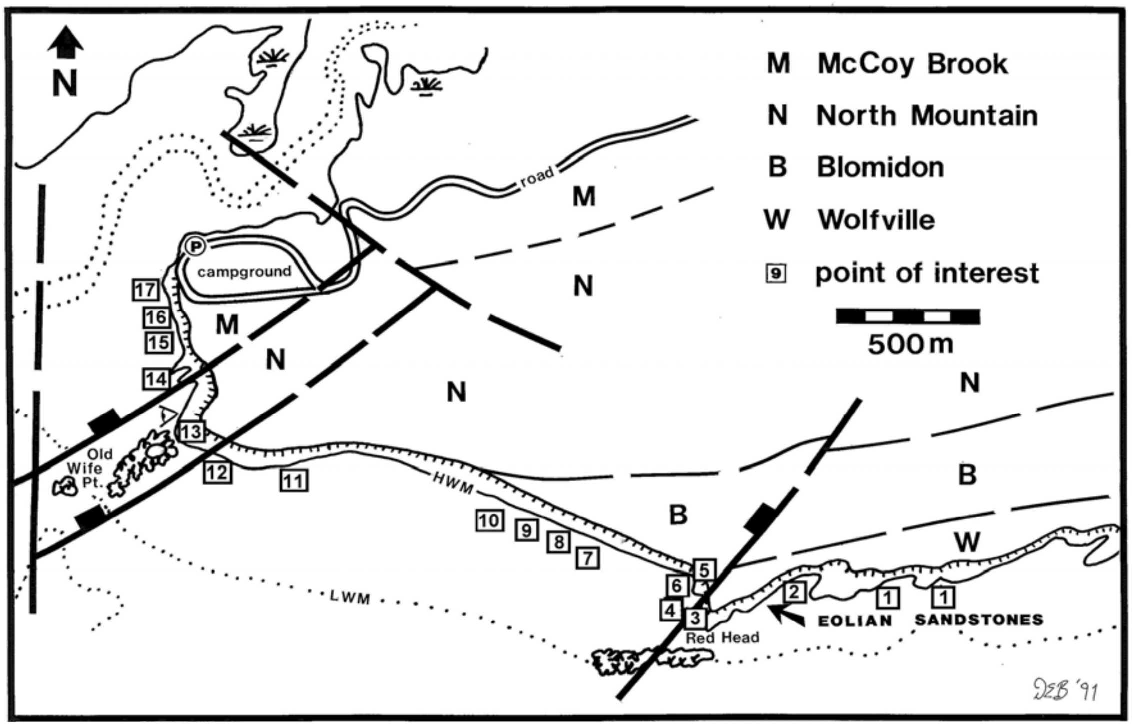


Figure 37: Geological map of the Five Islands Provincial Park area, Stop 3-1.

HIGHLIGHTS (in order of occurrence):
1. Stacked fluvial braidplain and alluvial fan channels (top Wolfville Fm.)
2. Aeolian dune field (base Blomidon Fm.)
3. Interdune facies and structures (base Blomidon Fm.)
4. Aeolian to playa facies change (base Blomidon Fm.)
5. Syndepositional (?) normal fault (aeolian and playa facies fault offset)
6. Domino-style, syndepositional faulting (Blomidon Fm.)
7. Playa/lacustrine cycles (Blomidon Fm.)
8. Individual sandflat cycles (Blomidon Fm.)
9. Lowstand muddy fluvial channel (Blomidon Fm.)
10. Thin sandpatch cycles (Blomidon Fm.)
11. Triassic-Jurassic boundary (Blomidon and North Mountain Fms.)
12. Fault contact (Blomidon and North Mountain Fms.)
13. Tectonized basalt flows (North Mountain Fm.)
14. Shallow-water muddy delta complex (McCoy Brook Fm.)
15. Overbank and levee deposits (McCoy Brook Fm.)
16. Rock fall / fish-kill bed (McCoy Brook Fm.)
17. Lacustrine and playa cycles (McCoy Brook Fm.)

Comments:
Trip participants will hike east along the shore with the receding tide to the end of the outcrop section where we will take some time to eat lunch. We will then work our way back up-section walking westward along the base of the cliffs towards the parking lot with the now in-coming tide.

The four formations encountered here at Five Islands are all in fault contact with each other, the near vertical faults being northeast-southwest trending and northwest-dipping. While exact displacements are uncertain, based on comparison with similar successions elsewhere in the basin they are thought to have 10s of metres displacements that significantly increase to the west. The entire succession dips gently between 15-20º to the north.

The excursion will begin about 100 m east of the large finger-like, east-west trending promontory about 200 m east of Red Head in the fluvial strata at the top of the Carnian Wolfville Formation. It will proceed into the thick aeolian dune sequence at the base of the Norian-Rhaetian Blomidon Formation and across a fault and then upwards through the Blomidon's playa-lacustrine succession. The Blomidon will be followed to the fault contact with the North Mountain Basalt at the craggy promontory known as Old Wife Point where one is afforded spectacular panoramic views of the outcrop section and region. The trip will cut across the tectonized basalt in this extensional setting and move over another normal fault into the muddy fluvial system of the overlying McCoy Brook Formation (Figure 37, Site 1).

Wolfville Formation

Figure 38: Photos of Wolfville Formation outcrops. **A)** Panoramic view to the east of Red Head showing sand-dominated stacked braidplain fluvial channels near the top of the Formation representing facies assemblage S2 of Fraley (1982). **B)** Close-up photograph of stacked planar and trough crossbedded fluvial sandtones capped by plane-bedded sandstones of facies assemblage S2 overlain by S3 aeolian sands (Fraley, 1982). Photo courtesy of Gela Crane (EnCana).

Exposures of near flat-lying red <u>fluvial sandstones</u> of the top of the Carnian Wolfville Formation are observed extending 5km east of Red Head to Economy Point forming a series of small coves and headlands (Figure 38A) (Site 1). The succession of amalgamated braidplain channel complexes has a structural dip slightly to the north and a dominant west-southwest palaeocurrent direction. Down-section to the east, the palaeocurrents in coarser alluvial fan sandstone and conglomeratic facies indicate a southerly direction. The flow direction of these sediments is south into the Minas Subbasin and then west towards the main Fundy Basin depocentre located near the western end of the Bay of Fundy (Figure 39).

The Wolfville strata on the northern margin of the Minas Sub-basin in the Red Head area is at least 250 m thick and record a complex series of first to fourth order, generally fining-upward coarse sand to conglomerate cycles reflecting the interplay between tectonism and deposition, and the sedimentary response to these processes (Fraley, 1988). Olsen (1997) places this succession within his tectonostratigraphic sequence TS-II representing initial synrift fluvial and aeolian sedimentation. The following description of this succession is based on Fraley's work unless otherwise indicated. Similar work on the Wolfville Formation on Sub-basin's south margin has done by Florenza (1982) and Hubert and Florenza (1988) and are representative of the TS-II early synrift sedimentation (Olsen, ibid.).

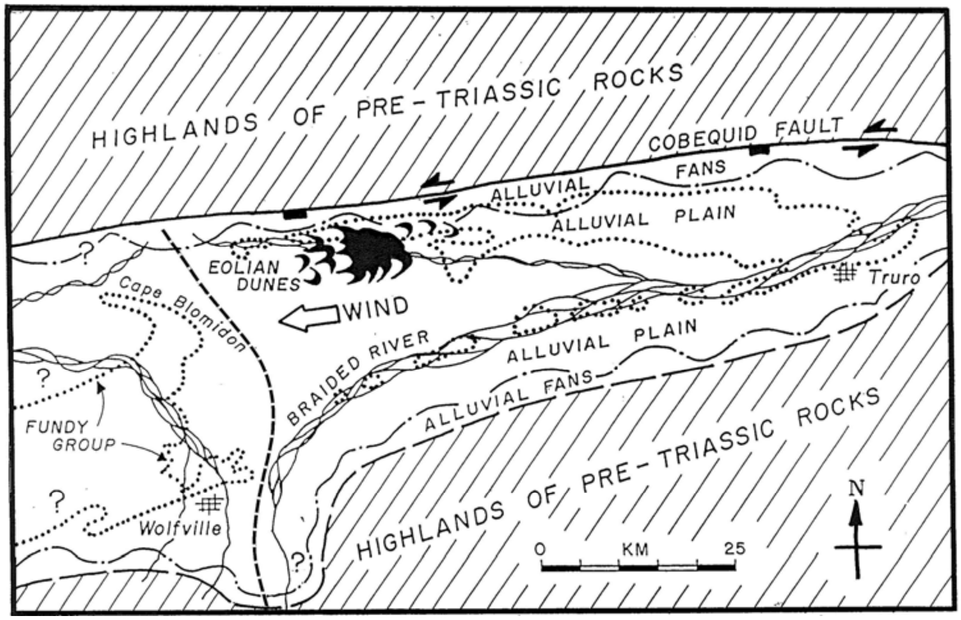


Figure 39: Palaeogeographic map of showing depositional facies during the Late Triassic (Wolfville) time. The present extent of outcrop of Fundy Group strata is within the areas outlined by the dotted pattern and the possible drainage divide by the dashed line (slightly modified after Hubert and Forlenza, 1988).

Several facies assemblages have been defined for Wolfville strata at this locality by Fraley (1982) based on study of the various facies developed within the Wolfville Formation:

(1). Horizontal bedded conglomerate and sandstone representing low relief to unchannelized high gradient sheetflood deposition as lobes on the distal portion of alluvial fans (Facies Assemblage 'G') ~35% of succession;

(2). Horizontally bedded, medium- to coarse-grained, occasionally pebbly sandstone and minor silt- and mudstone representing cyclic, ephemeral, high energy/high flow regime superimposed sheetflood facies deposited as lobes on the distal portion of alluvial fans (Facies Assemblage 'S1') ~30% of succession. This assemblage is the sandstone equivalent to assemblage 'G' (Figure 40A, B).

(3). Planar and trough cross-stratified sandstone representing high energy/low flow regime channel and bar complexes in a distal alluvial fan/proximal braidplain depositional environment transitional to the fluvial and basinal playa-lacustrine facies (Facies Assemblage 'S2') ~17% of succession (Figure 40C, D).

Subsidiary facies assemblages include aeolian sandstones ('S3'; ~16%) and basinal playa and playa-fringe mudstones ('F'; ~2%). Based on a comprehensive study of early Mesozoic rift sediments in the central Atlantic region and comparison with equivalent Moroccan successions, Olsen (1997) places the aeolian sequence at the base of the overlying Blomidon Formation.

Within the Wolfville succession at Red Head and further east to Economy Point, the defined facies assemblages have been used to identify various scale sedimentary cycles recording alluvial fan and braided stream deposition over time. These cycles reflect the interplay between tectonic activity and climatic conditions, and responsive depositional mechanisms to these factors such as fan lobe migration, fan head entrenchment, channel avulsion and evolution and changes in the source area. The entire Wolfville succession at this local defines a single, basin-filling first order cycle of about 500 m thick, within which are two asymmetric second order cycles (~120 m thick).

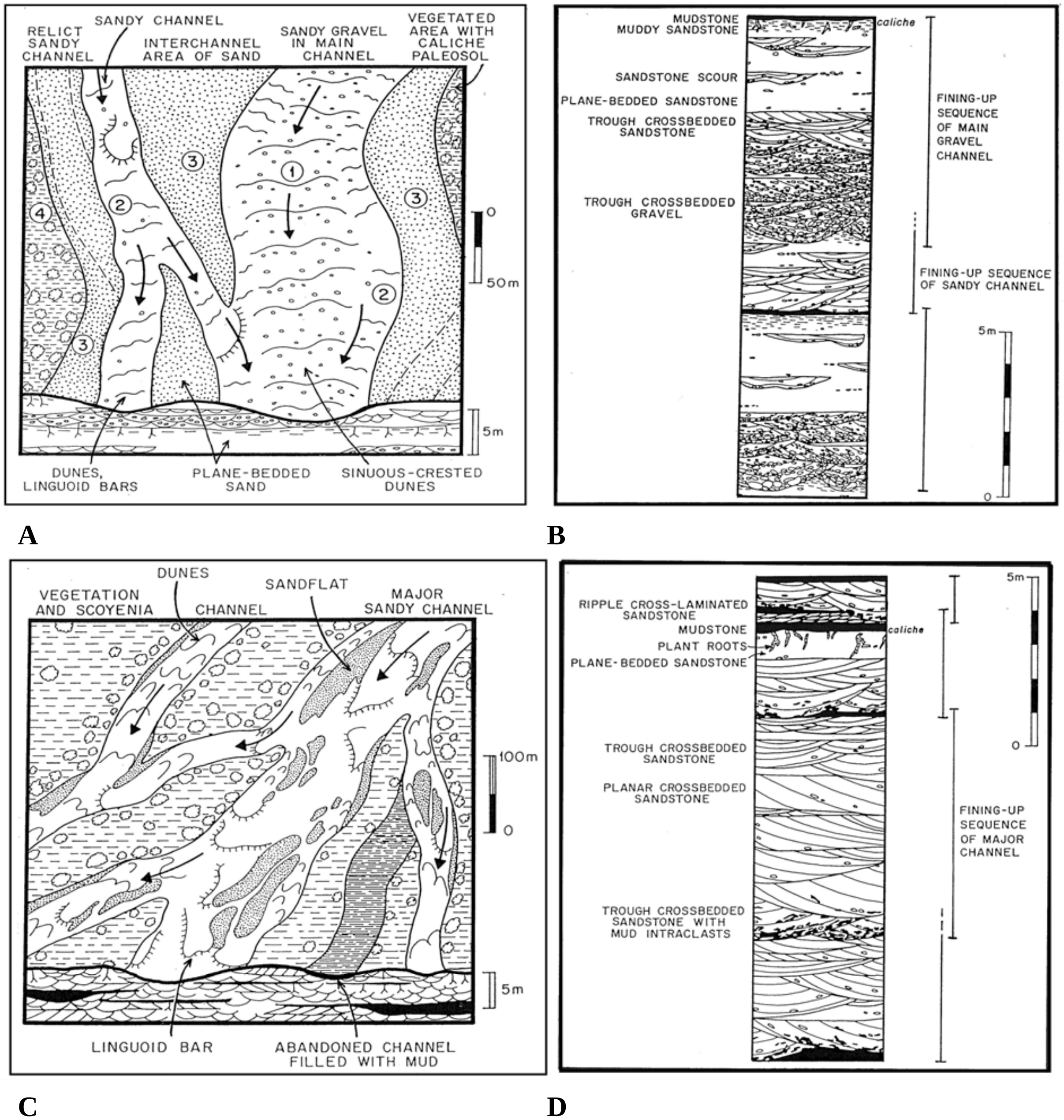


Figure 40: Wolfville Formation facies models. **A)** Model of a Wolfville Formation gravel-sand bedload river (Hubert and Forlenza, 1988), equivalent to the facies assemblage S1 of Fraley (1982). **B)** Typical set of fluvial, gravel-sand bedload, fining upward cycles in the Wolfville Formation, Burntcoat Head area on the southern margin of the Minas Subbasin (Hubert and Forlenza, 1988). Similar cycles are seen on the northern margin at Economy and are equivalent to facies assemblage S1 of Fraley (1982) **C)** Model of a Wolfville Formation sand-bedload river (Hubert and Forlenza, 1988). This is equivalent to the facies assemblage S2 of Fraley (1982). **D)** Typical set of fluvial, sand-bedload, fining upward cycles in the Wolfville Formation, also from the Burntcoat Head area (Hubert and Forlenza, 1988). Similar cycles are seen on the northern margin at Economy and are equivalent to facies assemblage S2 of Fraley (1982)

In this depositional setting, both cycles reflect primarily tectonic and climatic controlled deposition based on source area uplift and rejuvenation. The two distinct coarsening-thickening and fining-thinning second order cycles are interpreted based on facies assemblage 'G' clast size and thickness trends. Sand-dominated assemblages were found to be highly variable and thus less conclusive in defining cycles.

Within these two second order cycles are a number of third order fining-upward cycles (~15m thick). These record the response of sedimentary controls in the formation and evolution of alluvial fan depositional lobes such as channel avulsion, stream piracy and fan-head entrenchment. They represent deposition basinward of the intersection point where the upper fan head / trunk river trench ends and the active fan segment begins. These fining-upwards cycles are dominated by gravel dominated facies assemblage 'G' and subordinate sand-rich 'S1', 'S2' and 'S3' assemblages (Figure 38B).

Fourth order cycles (0.5m thick) are representative of individual flood events with deposition of horizontally stratified beds as the flooding subsides. These are commonly found in the distal fan lobe in the gravel ('G') and sand ('S1') dominant facies assemblages, especially in the former where they are manifested as planar cross-stratified gravel-sand wedge packages on the margins of channel longitudinal bars. In the sandier 'S2' facies assemblage with lower flow regime conditions, these beds are observed as planar cross-stratified transverse bars. Trough cross-stratified fining-upward beds in all assemblages record channel infilling by migrating mega-ripples during flood events.

At the base of the outcrop section at Lower Economy (not visited on this excursion), the alluvial fan sheet flood conglomerate and sandstone at the base of the Formation rest in angular unconformity on tilted lacustrine mudstone of the Mississippian Parrsboro Formation. The transition from fluvial to aeolian depositional environments at the top of the Wolfville Formation occurs over an approximately 3 metre thick zone of deflated fluvial sediments. The dunes are interpreted to have formed through aeolian reworking of fluvial and alluvial sediments at the distal end of alluvial fan complexes ('S3').

Blomidon Formation

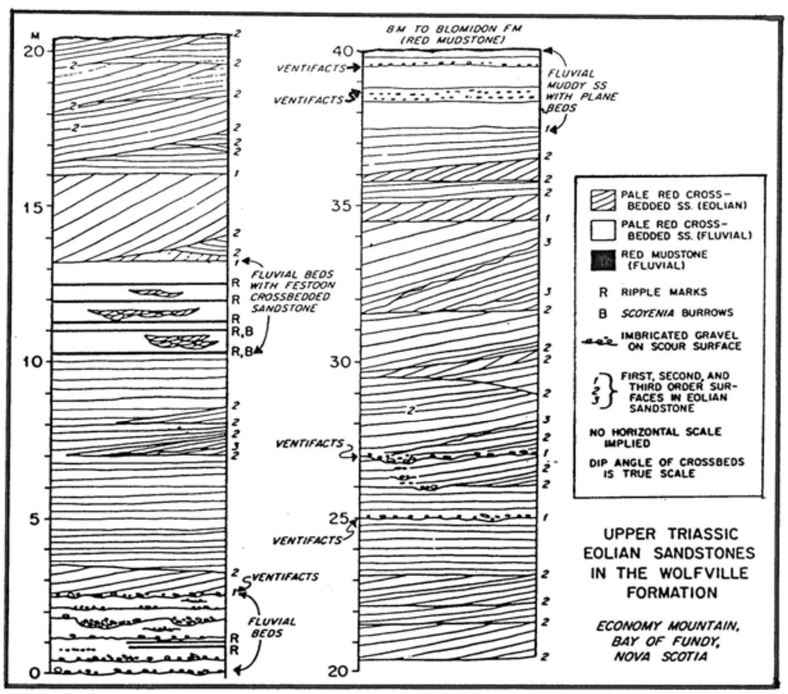


Figure 41: Total measured outcrop section of the basal Blomidon Formation ("Upper Wolfville") aeolian sandstones at Red Head, including bracketing fluvial muddy sandstones (Hubert and Mertz, 1980).

Spectacular exposures of Blomidon Formation aeolian dunes can be seen at Red Head and extending for several hundred meters to the east capping underlying braided stream sandstones (2). The total thickness of this sequence is about 33 metres, and was originally interpreted to be positioned at the top of the Wolfville Formation (Figure 41) (Hubert and Mertz, 1980, 1984). Recent work by Olsen (1997) places the dune sands at the base of the Blomidon Formation and records the initial onset of significantly drier climatic conditions and subsequent reworking of the fluvial sediments. Olsen (ibid.) places the Blomidon with his middle synrift tectonostratigraphic sequence TS-III and suggests a considerable hiatus is present between these apparently conformable formations.

A

B

C

Figure 42: Outcrop photos of the Blomidon Formation. **A)**. Strike section through complex transverse aeolian dunes showing first order (flat-lying), second order (low-angle crossbeds) and third order (internal tangential crossbeds) depositional surfaces. **B)**. Dip section through dunes showing wedge-planar and tangential crossbed sets bounded by flat-lying second order dune surfaces. **C)** Domino faults in playa-sandflat cycles at Red Head. The faults sole into a disturbed interval which itself extends into the main vertical fault a few metres to the right of the photos. Beds overlying this faulted interval remained undisturbed. Photos B & C courtesy of Gela Crane (EnCana).

Within the dune complex, a number of aeolian sedimentary structures can be observed in great detail (Figure 42A, B). The maximum thickness of an individual cross-bedded set is about 3 metres, averaging 1.6 metres (3). Minor fluvial (interdune) sandstone beds are also present. First-order erosional surfaces, which cross-cut all other aeolian features (second- and third-order surfaces), generally have a relief of 5-20 cm and in some cases display ventifacts and deflation-lag surfaces. The pebbles exposed on these surfaces consist almost entirely of early Palaeozoic age igneous rocks sourced from the adjacent Cobequid Highlands a few kilometres to the north. Second-order surfaces bound wedge-and/or tabular-planar, and tangential cross-bed sets which reflect changes in the paleo-wind direction and the resultant alteration in the direction of dune migration. Dune avalanche slip faces are representative third-order surfaces.

It is believed that this sequence represents an accumulation of compound transverse-type dunes with a palaeowind blowing from the northeast (Hubert and Mertz, 1980, 1984). At the top of the eastern wall of the small cove, there is a sharp conformable facies contact between the underlying aeolian and overlying playa-lacustrine sequences (4). This fourth order sequence boundary marks the top of the aeolian transitional interval between fluvial (wetter) and playa-lacustrine (drier) climatic conditions that dominate the Blomidon strata.

The aeolian sediments of the basal Blomidon are separated from its overlying playa-lacustrine strata by a near-vertical normal fault with an offset estimated at about 15 m (5). Alternating beds of orange sand flat sandstone and red lacustrine claystone/playa sandy mudstone reflect their lower position in the Blomidon section, in that they are dominated by the former lithology. Within the hanging wall, domino-style antithetic faulting is clearly exposed in these sediments immediately adjacent to, and soling into, this fault (6) (Figure 42C). These small faults have decimetre offsets yet show rotations exceeding 45º from the originally horizontal depositional surface, soling into a highly disturbed and brecciated red claystone layer. The faults climb stratigraphically away from the fault though laterally-intervening and overlying sediments are undisturbed confirming their syndepositional motion. Individual beds within the fault slivers can be traced laterally into undisturbed strata, with uninterrupted sedimentation eventually onlapping and burying the faulted sediments. It is believed that these features are the result of collapse from evaporite dissolution by groundwater triggered by seismic activity (Ackerman et al., 1995). This site thus displays beautifully the intimate and dynamic relationship between sedimentation and tectonics.

Almost 300 m of the Blomidon Formation playa-lacustrine sediments are spectacularly exposed for 1.5 km along the sea cliffs extending to the west from this site (7). Though the sediments of this formation generally record arid climatic conditions, the increasing frequency of thick red lacustrine claystone beds (and decreasing abundance of evaporite minerals in the sand flat/playa cycles) occurring beneath the basalt implies a return to wetter conditions (Olsen et al., 1989; Mertz and Hubert, 1990). Palynomorph data (Cornet, 1977) show the Triassic-Jurassic boundary to occur a few metres below the basalt within the ~2 m thick green-white interval (Olsen, ibid).

At the base of the section, above the aeolian sandstone, increasingly arid conditions are manifested by the presence of sand flat-playa cycles (8) (Figure 43). Detailed examination of these cycles is best undertaken in individual blocks in fresh rock slides that accumulate at the base of the cliff, though are quickly eroded and washed away. These are drier climatic variations of Olsen's lacustrine 'Van Houten cycles' (Olsen, 1985a, 1985b), and average about 1.5 metres thick and are laterally-continuous over the entire 1.5 km section. Each cycle usually begins with a thin (<1 metre) unit of red, occasionally mudcracked, laminated claystone and siltstone (Figure 44). Together with fish and other fossil remains, as well as oxygen and carbon isotopic data, a nearshore, shallow lacustrine environment is inferred (Mertz and Hubert, 1990).

The overlying sandstones consist of 5-30 cm thick fining-upward beds of horizontal, then ripple cross-laminated sandstone and then overlain by a thin mudstone drape. The sedimentary fabrics of these beds are commonly disrupted by the growth of gypsum and sometimes display evaporite crusts. The gypsum is ubiquitous as an interstitial filling in these sandstone beds, and as individual blebs. Minor beds and patches of aeolian sand are also encountered. Mertz and Hubert (ibid) interpret these sand units to have been laid down as sand flat deposits at the transition between distal, basin-margin alluvial fans and a basinal playa lake.

The culminating member in these cycles is a massive, disrupted, red sandy mudstone up to one metre in thickness. The unusual texture of the mudstone is known as sand-patch fabric (Figures 45A, 46) and is defined by Smoot and Olsen (1988) as "small (1-5 cm long), irregular pods of sandstone and siltstone having the following diagnostic characteristics: angular margins; internal, jagged, mud-filled cracks; internal zones of different grain sizes; and cuspate contacts with the surrounding mudstones", and "the pods may have internal cross-laminae that are not oriented with respect to cross-laminae in adjacent pods" (p. 261). At this location, the fabric is seen as buff-coloured patches of sand within red mudstone. The fabric is interpreted to represent the wind-blown and water-born deposition of silt and sand on a playa saline mud flat. The sediments settled into irregular depressions on the surface of muddy efflorescence evaporite crusts, and remained as a remnant lag within the mudstone following the dissolution of the evaporites (Smoot and Castens-Seidel, 1982). Clear gypsum spar is present as a cement/void-filling and as blebs within efflorescence crusts in the mudstone, though it is much more common in the coarser grain sand flat sandstone. This playa mudstone is the basin-centre equivalent to alluvial fan sheetflood sandstone and debris flow conglomerate.

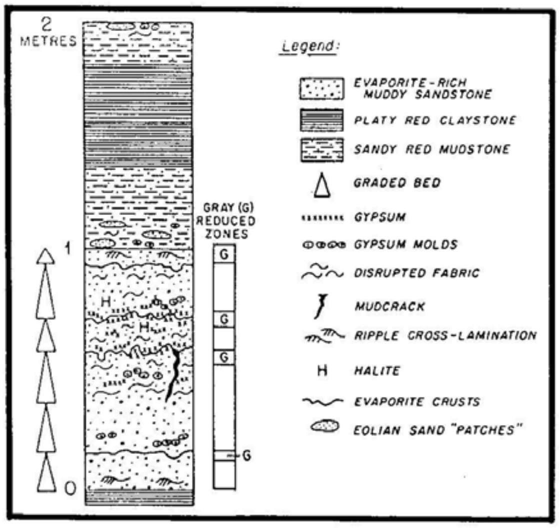


Figure 43: Generalized sand flat sandstone and playa-lacustrine mudstone cycle in the Blomidon Formation with the triangles indicating individual graded flood units (Mertz and Hubert, 1990).

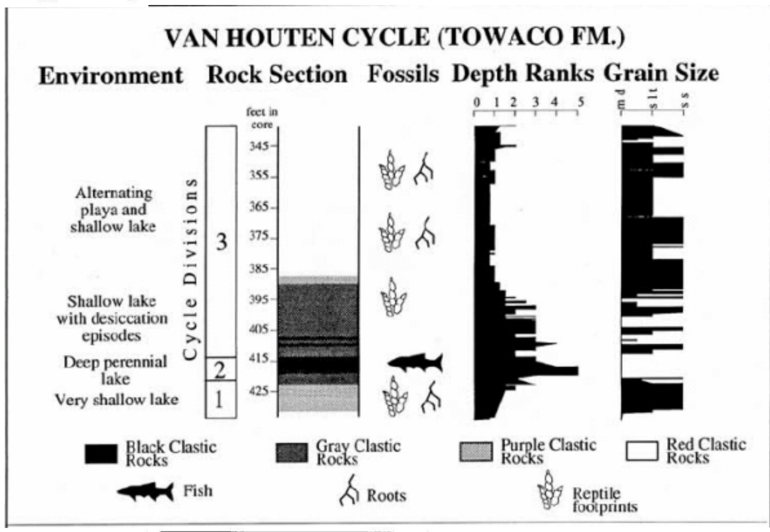


Figure 44: Model of a typical Van Houten Cycle (Olsen et al., 1996).

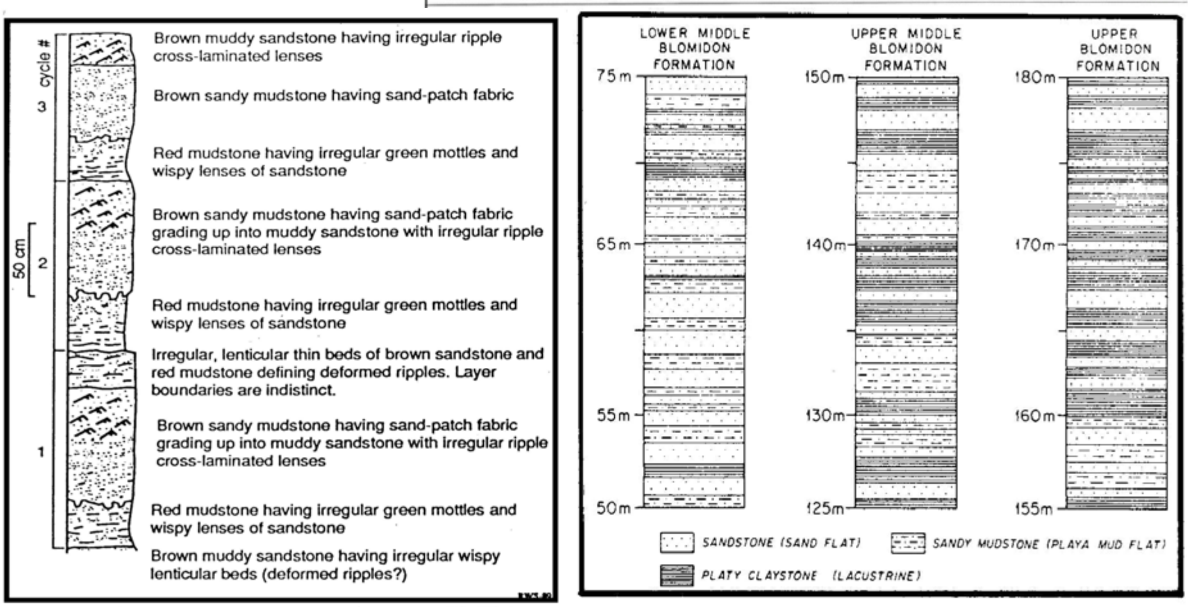


Figure 45: Sedimentological logs of the Blomidon Formation. **A)** A composite drawing of typical sand flat / playa-lacustrine sand-patch cycles (Olsen et al., 1989). **B)** Selected measured sections at Economy, illustrating the upward trend towards increasing occurrence and volume of lacustrine claystones and the simultaneous decrease in sand flat and playa mud flat strata (Mertz and Hubert, 1990).

Figure 46: Blomidon Formation sedimentology photos. **A)** Close-up photo of sandpatch fabric. **B)** Photograph illustrating sandpatch-playa/lacustrine cycles with thick red lacustrine clays reflecting increasingly wetter climatic conditions. Photos courtesy of Gela Crane (EnCana).

At the base of the cliff about 200 m west of the aeolian strata at Red Head is a metre-thick poorly sorted fluvial sandstone channel (9). In strike profile it gradually expands from its eastern end continuing westward for about 30 m until it is faulted out against playa sediments. This layer is interpreted as a fluvial channel filled with re-worked aeolian sandstone that in turn shows evidence of later aeolian deflation. The channel's palaeocurrent direction is southerly into the basin. Within the playa mudstone and siltstone under the channel's erosional base can be found large (5-8 cm) nodules of clear gypsum spar. Their presence infers preferential concentration of groundwater NaCl and $CaSO_4$ rich groundwater brines from which the nodules were precipitated.

Mertz and Hubert (1990) identified and measured over 100, 1-2 metre thick sand flat sandstone/playa-lacustrine sandy mudstone-claystone cycles in the Blomidon Formation (Figure 46B). These cycles are quite similar to those seen at Cape Blomidon, 35 km across the basin to the south, and can be traced laterally along the entire 1.5 km of exposed section here. The sandstone tends to form the more resistant beds, reflecting increased evaporite mineralization, mostly gypsum. The trend of these cycles over time, reflecting wetter climatic conditions, is revealed in an increase in the thickness, frequency, and total volume of red lacustrine claystone (a minor constituent within the playa mudstone at Cape Blomidon), along with a corresponding decrease in the above parameters for the sand flat sandstone and the playa mud flat sandy mudstone (Mertz and Hubert, ibid).

Thinner sandpatch cycles (20-30 cm) are dominated by deeper water lacustrine red to violet claystone beds and are encountered further west in the section (Figure 44B) (10). These may be studied in detail where rock falls are present. With a cumulative thickness of about 2 metres, these distinctive successions make excellent marker beds and from the vantage point at the Old Wife can be seen in the cliffs extending on to the east until truncated at the top of the sea cliffs. These 1-2 metre thick lacustrine-sand flat-playa cycles were postulated by Olsen (1989) to have been deposited over the 100,000 year orbitally-induced Milankovitch eccentricity cycle, and are found only within the Fundy Basin. Recent research (Olsen, 1997) suggests that these climatic precession cycles probably represent shorter 40,000 year obliquity cycles. The recognition of these facies, and that of aeolian strata at the base of the Formation (Hubert and Mertz, 1980, 1984) as well as the increasing occurrence of caliche paleosols within the lower Wolfville (Hubert and Forlenza, 1988), supports the interpretation of arid to later semiarid conditions existing during the Late Triassic (Cornet and Olsen, 1985) over the period of Blomidon deposition.

The Triassic-Jurassic boundary is beautifully exposed in the cliffs to the east, where the dark brown basalts overlie the bright orange-red Blomidon sediments (11) (Day 3 frontispiece). The 1-2 m thick light green-grey banded unit at the very top of the strata actually represents this boundary, and contains probable charred organic material (Olsen et al., 1989). The palynomorph assemblage in the top unit is dominated by the diagnostic Jurassic palynoflorule *Corallina meyeriana* (Cornet, 1977). Fossil evidence and the lack of observable hiatuses within the Blomidon Formation lead Mertz and Hubert (1990) to conclude that this Formation spans the entire Triassic Norian period and extends slightly into the Jurassic. They thus conclude the Blomidon has an age range of about 20 Ma, in which each one of the 100+ sand flat-playa/lacustrine cycles represent about 200,000 years of deposition, and with a sedimentation rate (factoring in compaction) of approximately 25,000 years per metre. Olsen (1997) however, suggests that these cycles are the products of only 40,000 years of sedimentation, which would significantly decrease the age span of the Blomidon Formation. The author favours the latter interpretation, given the evidence of dynamic sedimentation, ongoing subsidence and tectonism, and the 3000m+ thickness of Blomidon strata in the Fundy Basin (Wade et al., 1996).

North Mountain Formation

Continuing westward, gently north-dipping playa-lacustrine-playa sediments of the underlying Blomidon Formation are seen in <u>fault contact</u> with the basalts (12). The volcanics capping the Blomidon Formation were flat-lying, low viscosity, fissure-sourced continental tholeiite flood basalts. The North Mountain basalts have a U-Pb zircon age of 202±1 Ma (Hodych and Dunning, 1992), a date that matches very well indeed with the variously determined 201 Ma date for the Triassic/Jurassic boundary as reviewed by Olsen at al. (1989). They were probably extruded over a geologically brief period of time (i.e., ~580,000 years; Olsen et al., 1996; Puffer, 1990) and together these date it as earliest Jurassic (Hettangian).

Northeast-trending, down-to-the-west normal faults juxtapose the McCoy Brook Formation against the stratigraphically-lower basalts of the North Mountain Formation (Figures 47, 48). The <u>basalt breccia</u> at Old Wife Point exhibits near surface (?) <u>cataclastism</u> resulting from vertical faulting and shearing soon after deposition (13) (Stevens, 1980, 1987). Similar faults further west produced the prominent, craggy basaltic stack of the Old Wife. Looking to the NW from this point, a photogenic view is presented which well displays the stacked geometry of the deltaic channels in a section of about 35 m of McCoy Brook strata.

McCoy Brook Formation

From the vantage at Old Wife Point looking northwest, the McCoy Brook Formation is seen in fault contact with the underlying North Mountain basalts, with the 42 m thick exposure representing the top of the hanging wall (Figure 49A). In addition to this site, it is exposed in near continuous outcrop to Clarke Head 15 km to the west with about 230 m of strata measured. Over this interval, it displays a variety of terrestrial facies including talus breccia, alluvial fan, fluvial sandstone and mudstone, lacustrine shale and aeolian dune sandstone deposited directly on the surface of the North Mountain basalt (Tanner and Hubert, 1992). The complex facies distribution as observed in outcrop reflects the influence of the highly faulted basal depositional surface upon sedimentation. With eventual infilling and burial of this surface, the McCoy Brook strata appear relatively undisturbed and evenly blanket the basin margin. This succession represents the late synrift sedimentation of Olsen's (1997) tectonostratigraphic succession TS-IV.

The first section is a series of prominent, variably-thick, <u>alternating deltaic channel sandstone and overbank mudstone</u> (14) (Figure 49B). The channel sand deposits (Birney, 1985) are continuous over 100 m and range in thickness from 4.5 to 5.3 m. They are characterized by climbing, straight-crested and sinuous-crested ripples, cross- and flaser bedding, truncated foreset beds, and rip-up mud clasts, though the dominant lithofacies is horizontally laminated. The channels cut 20-40 cm into underlying overbank deposits and at their base are intraformational conglomerates composed of mud clasts that may contain fish bones and related coprolitic material (Olsen et al., 1989). These sandstone beds are interpreted to represent the results of ephemeral braided stream shallow flooding events in an upper flow regime setting (Tanner and Hubert, 1992). Large floods occasionally extended over the floodplain and deposited thin single story horizontally-laminated sandstone beds.

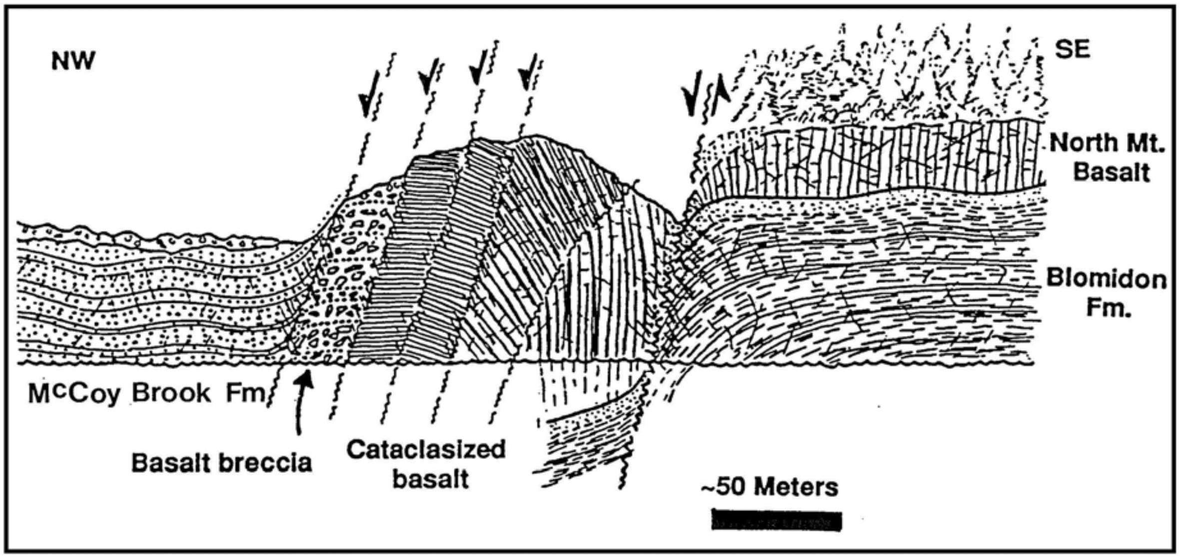


Figure 46: West-east cross section through Old Wife point illustrating the highly faulted and deformed North Mountain basalt separating the Blomidon and McCoy Brook formations (Stevens, 1980).

Figure 48: View to the northwest of the deformed North Mountain basalt at Old Wife Point. Light area to the right is undeformed columnar basalts. To the left, the basalts are folded, faulted and tectonized to create a 'ball-milled' texture of basalt clasts.

Surprisingly, palaeocurrent measurements indicate paleoflow directions to the north towards the basin margin highlands though is interpreted as evidence for a local slope on the floodplain and the response of sediment-laded floodwaters overspilling river channels (Tanner and Hubert, ibid.). The dominant lithofacies in this exposure are interbedded overbank/levee silty floodplain mudstone and sandstone (15) (Figure 49C). They are horizontally-laminated with occasional widely-spaced and occasionally very deep (~1m) desiccation cracks, root tubules and burrows and occasional palaeosols. Diagnostic Jurassic-age vertebrate footprints are also present in the mudstone (Olsen, ibid).

A modern talus slope / rock fall has transported to the shoreline large boulders of material from higher up in the channel/overbank sequence (16). In fresh falls, the boulders are sometimes composed of a red, cross-laminated sandstone overlying a violet-red laminated claystone. At this contact, whitish, densely-packed, complete fish fossils (*Semionotus* sp.) are common and it would appear that this interval is representative of a single fish-kill event (Olsen et al., 1989).

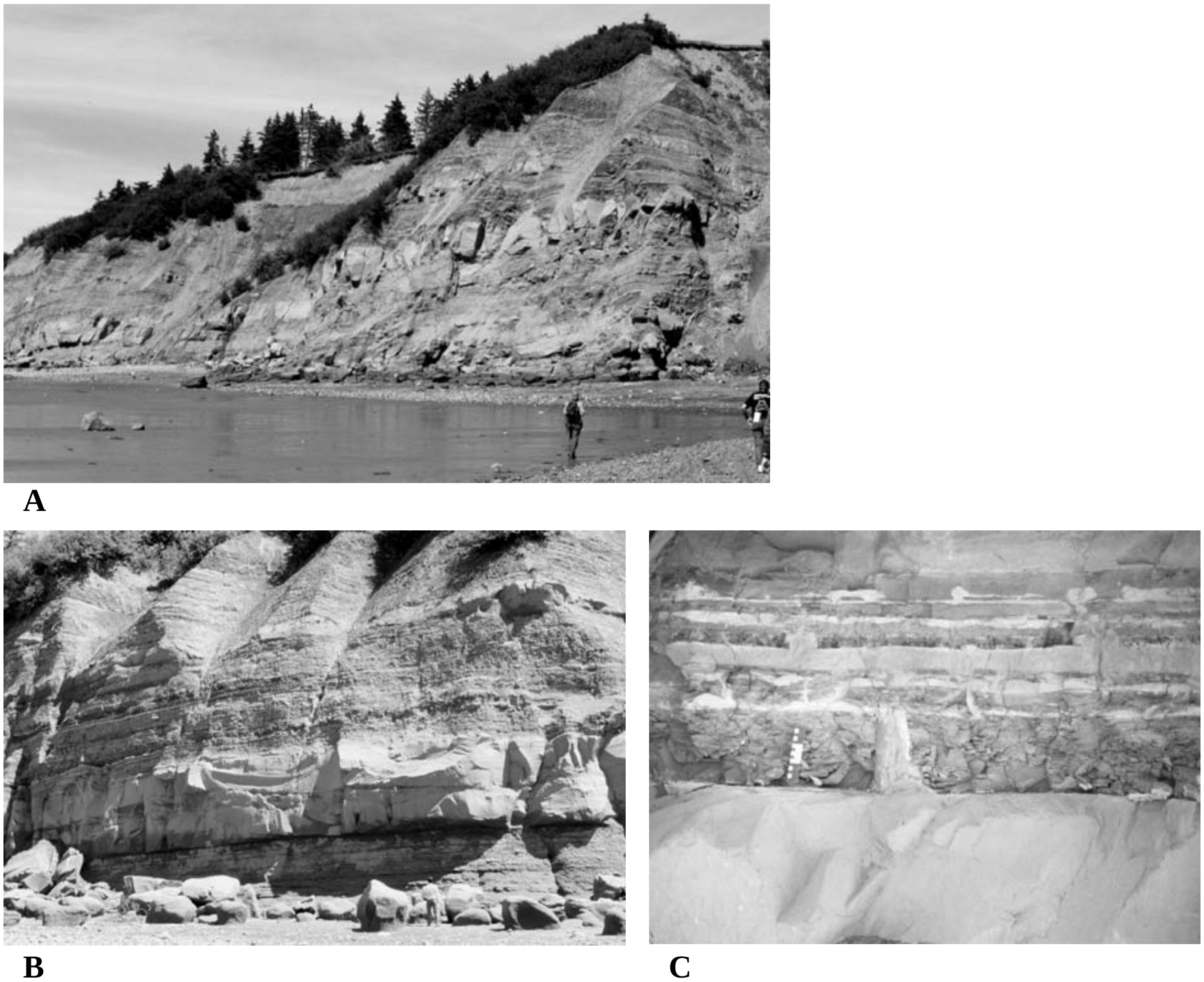


Figure 49: Photographs of outcrop of the McCoy Brook Formation. **A)** Northwest view from Old Wife Point of the stacked fluvial channels. **B)** Close up of one of the three main channel complexes. Amalgamated channels are separated by third order surfaces with a thin planar-bedded sand bed capping the sandstone. The channel is encased in overbank and levee silty mudstones. Playa and lake margin sediments cap the outcrop. Note that there is virtually no evidence of incision and relief at the base of the channels. Semionotid fish remains can sometimes be found in fresh boulders from occasional rock falls that accumulate in the area to the left of the image. **C)** Close-up of sand-filled mud cracks in marginal lacustrine/overbank-levee overbank mudstones overlying a channel interval.

The last McCoy Brook strata encountered are reddish-brown marginal-lacustrine mudstone and sandflat sandstone (17). Desiccation features (prism cracks, sand-patch fabrics, etc.) are not that common in these rocks, suggesting a reduction in the frequency of total lake desiccation. This may be coupled with increased precipitation rates that reflect the previously discussed trend of decreasing aridity from the latest Triassic on into the Jurassic.

The age range for the McCoy Brook Formation is uncertain. The variety of palaeontological and radiometric analyses and extrusion models indicate that at the base of the Formation, the underlying North Mountain basalt is earliest Hettangian in age and therefore deposition can be dated as initiating at this time. Macrofossil (Olsen, 1981; 1988) and palynoflora data above the basalt from outcrops and offshore wells confirm a Hettangian age. The appearance of Late Triassic palynomorphs near the base of the Formation suggests probable local erosion and reworking of disturbed Blomidon Formation or equivalents along the northern faulted margin of the Fundy basin.

The upper age limit of the McCoy Brook Formation, and hence the entire Fundy Basin succession, can only be estimated. Traverse (1987) reported evidence of strata as young as Pliensbachian in the Fundy Basin and Olsen (1988) indicated the presence of Pliensbachian-Toarcian sediments in the Economy/Parrsboro area of the Minas Sub-basin. Offshore, seismic data indicates that about 3000 metres of Scots Bay Formation strata are present in the Fundy Basin depocentre near Grand Manan Island and probably about 2000 m or more of additional Scots Bay or younger strata removed by later erosion (Wade et al., 1996).

End of Day 3

En Route

Return to Route 2, turn right and head east ~45 km to Glenholme and the junction with Route 4. Turn right and follow the road 6 km to the junction with Highway 104 at Masstown. Follow the highway ~15 km east to Exit 15 (ONLY exit to Halifax/Truro!). Veer right onto Highway 102 and head south, passing the town of Truro towards Halifax. After about 40 minutes driving and passing Exit 7 to the Halifax International Airport, continue on ~15 km to Exit 5 (Millers Lake of left). Keep in the left lane at Exit 5 onto Highway 118 and drive about 13 km to Exit 4. Veer right and onto Highway 111 drive 3.5km to the A. Murray McKay Bridge (toll = $0.75), cross the bridge and follow the signs to the Windsor St./ Bedford Hwy intersection (centre lane - ~1km from end of bridge). Follow Windsor Street about 300 m up the hill (cemetery to the right) and at the corner intersection lights turn right onto Connaught Avenue. Follow this avenue 2.7 km to its end at Jubilee Road. Turn left at the 3-way stop intersection and proceed 200 m to the next set of lights at Oxford Street. Turn right and drive 300 m to the next set of lights at Coburg Road. Turn left and continue about 200 m and turn right onto LeMarchant Street. Drive about 100 m to the intersection with University Avenue. Ahead to the left and on the corner is the Dalhousie University Graduate House. Cross the boulevard and turn left. Park on University. End of trip.

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Pre-conference Field Trips

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

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

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

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

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

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

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

Post-conference Field Trips

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

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

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

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

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

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

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

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

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