GEOSCIENCE CANADA

JOURNAL OF THE GEOLOGICAL ASSOCIATION OF CANADA
JOURNAL DE L'ASSOCIATION GÉOLOGIQUE DU CANADA

Gros Morne National Park – Outings in its Ancient Oceanic Crust

The Bathurst Mining Camp – A Canadian Giant Revisited

The Reign of Dinosaurs – From Triassic Beginnings to the K–T Boundary

2019 VOLUME 46

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Print Edition: ISSN 0315-0941
Online Edition: ISSN 1911-4850

Volume 46

A journal published quarterly by the Geological Association of Canada, incorporating the Proceedings.
Une revue trimestrielle publiée par l'Association géologique du Canada et qui en diffuse les actes.

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Cover Image: The Moho in Gros Morne National Park, western Newfoundland; the grey rocks to the left are gabbro of the oceanic crust, and dark brown rocks to the right are mafic and ultramafic cumulates transitional into the upper mantle. Photo credit: John Waldron.

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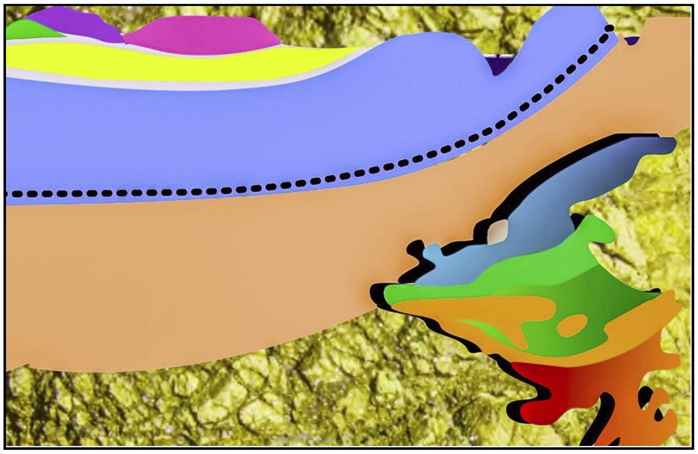

Classic Rock Tours 2.
Exploring a Famous Ophiolite: A Guide to the Bay of Islands Igneous Complex in Gros Morne National Park, Western Newfoundland, Canada

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SUMMARY

Ophiolites are complex assemblages of ultramafic and mafic igneous rocks that are now widely considered to be pieces of ancient oceanic crust that were emplaced on to the continents courtesy of global plate tectonics. However, most examples were originally considered parts of enormous layered mafic intrusions and so were interpreted in that light. The new understanding of ophiolites in the late 1960s and early 1970s was a crucial part of the global Earth Science revolution, and they are now central to all plate tectonic models developed for ancient orogenic belts. Although their equivalence to oceanic crust is now well established, many ophiolites may not be 'typical' examples of such, and not all examples are identical. Most ophiolites likely formed in subduction-influenced environments rather than at mid-ocean ridges. Ophiolites remain important foci for research in the 21st century, and many questions remain about their environments of formation and especially their mechanisms of emplacement onto the continents.

Although it was not the first to be seen as a relic of a vanished ocean, the Bay of Islands Igneous Complex in western Newfoundland is one of the best preserved and most easily accessible ophiolites in the world. In the late 20th century, research work in this area proved highly influential in understanding the oceanic crust, and in unravelling the diachronous events involved in the progressive destruction of an ancient stable continental margin as arcs and microcontinental blocks were accreted along it. Parts of the Tablelands Ophiolite lie within Gros Morne National Park, which is a UNESCO world heritage site because of its importance to our understanding of global tectonics. The wider region around the park also includes the Cabox Aspiring Geopark Project, now also in the process of seeking recognition through UNESCO.

This article provides background information on ophiolites and the development of our ideas about them, and links this material to four self-guided field excursions that allow examination of many classic features. These excursions range from a collection of roadside outcrops, to some relatively easy hiking excursions on official National Park trails, and eventually to a more challenging off-trail hike that ascends to the summit plateau of the Tablelands to visit rare exposures of the Moho (the Mohorovičić Discontinuity, i.e. the lower boundary of the Earth's crust) and the underlying upper mantle rocks. Collectively, the field stops should allow geologically-minded visitors to experience some amazing geology in a spectacular and sometimes surreal landscape.

RÉSUMÉ

Les ophiolites sont des assemblages complexes de roches ignées ultramafiques et mafiques qui sont maintenant généralement considérées comme des fragments de croûte océanique ancienne qui ont été charriés sur les continents grâce à la tectonique globale des plaques. Cependant, la plupart des exemples étaient à l'origine considérés comme faisant partie de vastes intrusions mafiques stratifiées et ont donc été interprétés dans ce contexte. La nouvelle compréhension des ophiolites à la fin des années 60 et au début des années 70 a été un élément crucial de la révolution des sciences de la Terre. Les ophiolites sont désormais au cœur de tous les modèles tectoniques des plaques développés pour les anciennes ceintures orogéniques. Bien que leur équivalence avec la croûte océanique soit maintenant bien établie, de nombreuses ophio-

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Geoscience Canada, v. 46, http://www.dx.doi.org/10.12789/geocanj.2019.46.149 pages 101–136

lites peuvent ne pas en être des exemples « typiques », et tous les exemples ne sont pas identiques. La plupart des ophiolites se sont probablement formées dans des environnements influencés par la subduction plutôt qu'au niveau des dorsales océaniques. Les ophiolites restent un thème de recherche important au XXIe siècle et de nombreuses questions subsistent quant à leurs environnements de formation et notamment à leurs mécanismes de mise en place sur les continents.
Bien qu'il n'ait pas été le premier à être identifié comme un vestige d'un océan disparu, le complexe igné de la baie des Îles, dans l'ouest de Terre-Neuve, fait partie des ophiolites les mieux conservées et les plus facilement accessibles au monde. À la fin du XXe siècle, les travaux de recherche dans ce domaine ont joué un rôle déterminant dans la compréhension de la croûte océanique et dans la compréhension des événements diachrones impliqués dans la destruction progressive d'une ancienne marge continentale stable au fur et à mesure de l'accrétion d'arcs et de blocs microcontinentaux. Une partie des Tablelands Ophiolite se trouve dans le parc national du Gros-Morne, site classé au patrimoine mondial de l'UNESCO en raison de son importance pour notre compréhension de la tectonique globale. La région plus large autour du parc comprend également le projet Cabox Aspiring Geopark, qui est également à la recherche d'une reconnaissance dans le cadre de l'UNESCO.

Cet article fournit des informations de base sur les ophiolites et le développement de nos idées à leur sujet, et relie ce matériel à quatre excursions autoguidées qui permettent d'examiner de nombreuses caractéristiques classiques. Ces excursions vont d'une collection de visites d'affleurements au bord de la route, à des randonnées relativement faciles sur les sentiers officiels du parc national, et finalement à une randonnée plus difficile hors-piste menant au plateau sommital des Tablelands pour visiter de rares affleurement du Moho (la discontinuité de Mohorovičić, c'est-à-dire la limite inférieure de la croûte terrestre) et des roches du manteau supérieur sous-jacentes. Collectivement, les visites sur le terrain devraient permettre aux visiteurs amateurs de géologie de faire l'expérience d'une géologie remarquable dans un paysage spectaculaire et parfois surréaliste.

Traduit par la Traductrice

PREAMBLE

In the late 1960s, ideas long held by geologists about stratigraphy, mountain belts, the origins of igneous rocks and many other concepts were revised and in some cases entirely reversed. Unusual mafic and ultramafic 'intrusions', which had long puzzled both mapmakers and petrologists, assumed greater importance when it was first proposed that they might be remnants of the vanished oceans that global tectonics predicted. The term *ophiolite* is now almost as familiar as 'basalt' to undergraduates, but few of them realize that its use long predates ideas of continental drift, or that these enigmatic rocks took so long to understand. Most of the ophiolites around the world were once studied largely by petrologists and mineralogists. These vast igneous massifs were logically interpreted as large layered intrusions that were younger than the surrounding rocks, even if the contact relationships were enigmatic. The evidence for the true nature of ophiolites was, of course, always there for us to record but it was discounted because the implications extended beyond prevailing concepts. Such is often the case in the history of geology.

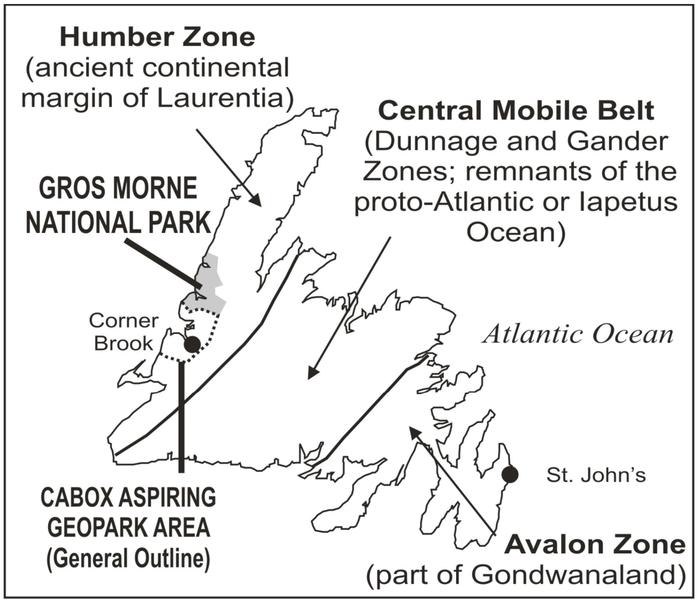

Figure 1. The location of Gros Morne National Park and the adjoining Cabox Aspiring Geopark areas in western Newfoundland. The map also shows the fundamental three-fold division of Newfoundland into the Humber Zone (the ancient continental margin of Laurentia), the Avalon Zone (Late Precambrian rocks that correlate with similar rocks in Europe and North Africa) and the Central Mobile Belt (Dunnage and Gander Zones, which represents island arcs and other regions formed within the proto-Atlantic or Iapetus Ocean). Terminology for the Appalachian Orogen in Newfoundland after Williams (1979, 1995).

There are many well-established (but variably understood) ophiolites around the world, and the list continues to expand (e.g. Dilek and Furnes 2011, 2014). Ophiolites are key elements in our interpretations of ancient orogenic belts and the development of tectonic models for them (e.g. Dalziel et al. 2000). The ultramafic rocks within ophiolites commonly form unusual and striking landscapes, so it is not surprising that several important examples are within National Parks or other reserves. Gros Morne National Park in western Newfoundland (Fig. 1) includes parts of the Cambrian–Ordovician Bay of Islands Igneous Complex (BOIC), which was first recognized as an ophiolite in the late 1960s. Western Newfoundland is a key location where ideas linking ancient rocks and the modern oceanic crust were first integrated (e.g. Stevens 1965; Church and Stevens 1971; Williams and Stevens 1974). The BOIC was not the first ophiolite to be interpreted in this way, but it has proved to be one of the most influential, and it is certainly one of the most spectacular. The ophiolites are one of the primary reasons why Gros Morne National Park is now designated as a UNESCO World Heritage Site. Beyond the National Park boundaries, the BOIC also forms the core of the Cabox 'Aspiring Geopark' proposal, also under the broad umbrella of UNESCO (see www.caboxgeopark.org for more information). Ophiolite enthusiasts undoubtedly have their own favourites (e.g. Bédard 2014), but most would agree that western Newfoundland is one of the best places on Earth to explore an ophiolite and understand the story that it tells. This article is intended to help make that process easier for geologically-minded visitors to this now-famous area.

This article does not pretend to be a specialized research contribution, but instead sets out to review background information and safely guide those who wish to independently explore and appreciate this influential 'type area'. Much of the material herein is derived from previous field trip guides and articles of a more specialized nature (e.g. Malpas 1987; Calon et al. 1988; Williams and Cawood 1989; Berger et al. 1992; Cawood and Suhr 1992; Stevens et al. 2003). Additional information, and most photographs, are from many personal visits and a long-dormant project with the late Robert K. Stevens to write a more comprehensive geological guidebook to the National Park. The numerous publications referenced in the text should remain the primary citations for details of the geology, rather than this more generalized overview. The article is intended to be understood by readers who lack specialized knowledge of tectonic processes or igneous rocks, but it assumes a general understanding of geological terminology and concepts appropriate to introductory Earth Science courses. A short glossary is provided in the appendix to explain specific terms that may be unfamiliar, and one of the many published dictionaries of geology might also prove useful. For those with minimal knowledge of geology, the excellent book by Hild (2012) covers parts of Gros Morne, including the Tablelands and provides information intended for a wider non-scientific audience.

OPHIOLITE ESSENTIALS

The Development of the Concept

The term *ophiolite* long predates the plate tectonics revolution. It is derived by combining *ophio-* (Greek for 'snake') and *'-lite'* (a Greek suffix for 'stone'), so its original meaning is *snakestone.* This curious name reflects the abundance of serpentinites - as you might guess, serpentine minerals were so named because their weathering textures resemble snakeskins. Ophiolite was an early descriptive term used for mafic and ultramafic rocks, commonly altered and variably deformed, that were first described within the Italian Alps (Brongniart 1821). The later term *Steinmann Trinity* captured the common association of such serpentinites with altered basalts ('spilites') and cherts - a threesome usually found together somewhere within ancient orogenic belts. Steinmann (1927) noted the deep-water character of the laminated cherts, and implied a link between these three rock types and the deep ocean floor. Interestingly, the area discussed by Steinmann is a rather atypical ophiolite, as it is missing some characteristic components (e.g. Bernoulli 2001; Desmurs et al. 2001). For over a century, ophiolites were interpreted as large intrusions, but they invariably had ambiguous relationships with surrounding sedimentary and volcanic rocks. The lack of large contact metamorphic aureoles around them seemed inconsistent with high-temperature mafic or ultramafic magmas, and it was hard to confirm that they intruded their country rocks. Also, prevailing ideas about igneous processes (e.g. Wager and Deer 1939) suggested that such ultramafic rocks had to be derived by processes of crystallization and mineral accumulation within much larger mafic magma chambers, which never seemed to be preserved in the vicinity. They were truly a puzzle.

The idea that ophiolites might be *tectonically* emplaced in a solid state was suggested first for the California Coast Ranges, and then for similar bodies in the South Island of New Zealand (e.g. Coleman 1967, 1971). However, the idea that they might originate elsewhere did not solve the puzzle, because it was not clear where 'elsewhere' might be. Prior to the oceanic drilling and geophysical programs of the 1960s, geologists had limited knowledge about oceanic crust and little inkling of the many ways in which continents and oceans could interact. A new interpretation of ophiolites emerged in the light of clearer evidence for the mobility of continents and the transience of oceans - they were seen as slices of oceanic crust and underlying mantle. Although the first such musings came from California, interpretation of the Troodos Mountains (Cyprus) as uplifted oceanic crust (Gass 1968) first brought the idea into the scientific mainstream. Coleman (1971) first coined the term *obduction* for the process by which ophiolites are tectonically emplaced, but the exact details of this process are still open to debate, as discussed in a later section.

With new interpretations of Mesozoic ophiolites in hand, it was logical to think that Paleozoic examples might also be relics of long-vanished oceans. If so, ophiolites could provide key evidence for the operation of global plate tectonics through Earth's history. The Bay of Islands Igneous Complex in western Newfoundland (Fig. 1) was one of the first to be reinterpreted (Stevens 1965; Church and Stevens 1971; Dewey and Bird 1971), and it remains a classic area. Western Newfoundland (and Gros Morne National Park in particular) has wider importance in plate tectonics, because the rocks reveal how various components of an ancient continental margin were transported and juxtaposed as island arcs, and microcontinental blocks were progressively 'accreted' to North America from the adjoining Iapetus Ocean (see later discussion for the origins of this name). The spectacular ultramafic rocks that symbolize Gros Morne National Park are the best-known part of this story, but they are only one chapter in a much longer tale.

To say that ophiolites are critical to understanding Earth history is a huge understatement. Almost all Paleozoic orogenic belts contain ophiolites, although their preservation varies widely, but in all cases they constrain plate-tectonic models. Ophiolites of Precambrian age remain elusive, but several are now widely accepted, including the well-preserved Paleoproterozoic Purtiniq ophiolite of northern Quebec (e.g. Scott et al. 1992). Ophiolites remain important research topics today, and there is even a scientific journal completely devoted to their study (*Ofioliti*, published in Italy; www.ofioliti.it). My general article and guide simply cannot do justice to decades of study and many remaining questions. Several papers in the journal *Elements* (2014, volume 10, issue 2) provide reviews with extensive bibliographies. An earlier GSA Special Paper

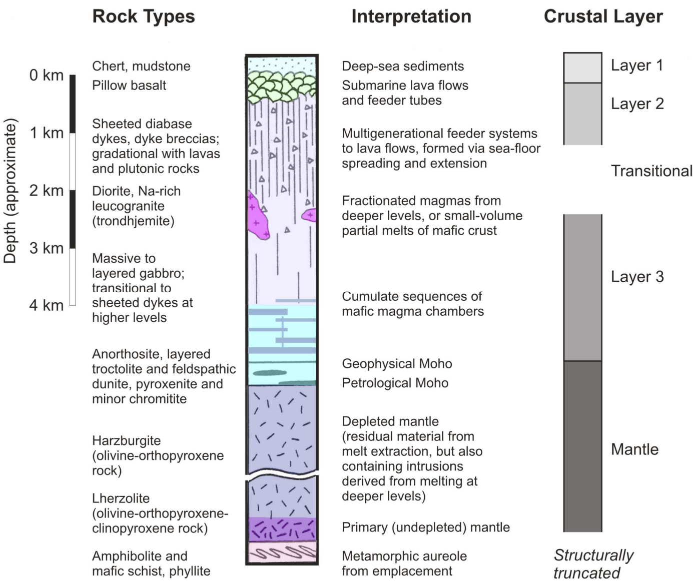

Figure 2. The major components of the ophiolite suite, their interpretation and their relationship to layer structure defined in the modern oceanic crust by seismic and other geophysical methods. Modified after Williams and Malpas (1972) with specific reference to the Bay of Islands Igneous Complex, but also applicable elsewhere. More detailed views of various types of ophiolites are provided by Dilek et al. (2000) and Dilek and Furnes (2014).

(Dilek et al. 2000) provides a wealth of more detailed information on specific examples. The following sections summarize some key aspects, with specific application to the western Newfoundland examples highlighted in this article.

Components of Ophiolites and Links to Modern Oceanic Crust

Although terminology varies, seven principal components are recognized within ophiolites and equated with specific parts of the oceanic crust (Fig. 2 after Williams and Malpas 1972; Dilek and Furnes 2011, 2014), even though our *direct* knowledge of the latter is limited to the uppermost 2 km or so. From top to bottom, these are:

1. Sedimentary rocks (typically cherts or shales) that were deposited on the lavas forming the deep-seafloor.
2. Pillow lavas of basaltic composition that were extruded to form the seafloor.
3. 'Sheeted dykes' of basaltic (diabase) composition, which are interpreted as the feeder systems to the overlying pillow lavas.
4. Gabbroic rocks representing high-level fractionated magma chambers, commonly gradational with the overlying sheeted diabase dyke systems.
5. Layered gabbros, which record cumulate processes in the deeper sections of mid-crustal magma chambers.
6. Ultramafic cumulate rocks, notably feldspathic dunite and olivine-rich gabbro, which define the transition between crust and mantle, i.e. the Moho, which is shorthand for ‘Mohorovičić Discontinuity’.
7. Massive (but commonly deformed) ultramafic rocks (harzburgite, dunite and pyroxenite) that represent the residual mantle material from partial melting that produced the overlying oceanic crust section.

Geophysical studies of the modern oceanic crust indicate that it is layered, and four layers are generally defined on the basis of its seismic velocity structure. The seven components listed above for ophiolites are broadly equated with these seismically-defined layers (Layers 1 to 4; Fig. 2), but they do not necessarily have individual seismic expression. The original definitions suggested from a landmark Penrose conference (Penrose Field Conference Participants 1972) excluded the uppermost sedimentary rocks, and it is now recognized that modern oceanic crust may lack parts of layer 2 (pillow basalts and sheeted dykes) and layer 3 (gabbro) in areas of strong extension. Such is apparently the case for the ophiolitic rocks that first defined Steinmann’s trinity (Desmurs et al. 2001). Ancient ophiolites are rarely complete, as they may be structurally dismembered during emplacement and/or truncated by later erosion. The dyke- and basalt-dominated upper parts of the ophiolite assemblage are the most vulnerable to removal by both mechanisms, and may be absent. Ultramafic rocks representing the upper mantle are more robust, although they may be variably serpentinized. Subsea hydrothermal systems cause extensive alteration, and emplacement involves low-grade metamorphism, so the original features of many ancient ophiolites are cryptic.

Although not specifically described above, many ophiolites also contain minor quantities of granitoid rocks, which are typically plagioclase-rich (albite) tonalites, also known as *trondhjemites*. These are interpreted as residual magmas from deeper fractionating magma chambers, or as products of localized partial melting within the crust (e.g. Malpas 1979; Jenner et al. 1991; Dilek and Furnes 2014). Intermediate and felsic magmas of trondhjemitic type are also known in modern arc-type settings (Dilek and Furnes 2014), and their presence in ancient ophiolites is important in understanding their tectonic settings.

Some ophiolite complexes, including the BOIC, also contain discontinuous belts of metamorphic rocks that are located structurally *below* the ultramafic rocks. These are commonly of mafic composition, and are interpreted as thermal and tectonic aureoles developed during obduction. These metamorphic aureoles provide very important evidence for the tectonic emplacement of ophiolites and their high ambient temperatures compared to structurally underlying rocks. The metamorphic aureole of the BOIC was one of the first to be recognized and interpreted in this fashion (Williams and Smyth 1973).

Ophiolites host specific styles of economic mineralization although these are not in themselves diagnostic features (e.g. Jébrak and Marcoux 2015). Examples include sea-floor sulphide deposits (volcanogenic massive sulphides, or VMS) in volcanic rocks, and podiform to massive chromite accumulations in peridotites. Chromite occurrences are typically small, but many ophiolites were also explored for stratiform chromitites and platinum-group element deposits because the mafic and ultramafic rocks were considered to represent parts of large layered mafic intrusions, which host important deposits of both types. Important asbestos deposits occur in many ophiolites, formed by syn-and post-emplacement alteration of the ultramafic rocks; notable Canadian examples include the Baie Verte area in Newfoundland and the Thetford Mines area in Quebec (Jébrak and Marcoux 2015). Diamonds are reported from some ophiolite complexes (e.g. Yang et al. 2014), although such concentrations are not thought to be economically significant.

Many studies over many years have outlined the strong similarities between ancient ophiolites and the modern oceanic crust in terms of rock types, composition and structure. The comparison is not perfect in all details for all examples, but the discrepancies lie mostly in the relative abundance of rock types, or petrological details, rather than gross anatomy. Our *direct* knowledge is probably more complete for ancient ophiolites than for the modern oceanic crust, and we cannot be certain that original relative thicknesses are retained following their obduction. More importantly, many ophiolite sequences probably represent substrates to island arcs rather than true mid-ocean-ridge settings (see later discussion) and such arc environments are intrinsically more varied in character. Notwithstanding continuing discussion as to exactly *where* individual ancient ophiolites were formed and *how* they were emplaced, the link to the crust of the modern oceans remains convincing, and such a conclusion is also supported by the similarities of seismic profiles obtained from the BOIC (and other ophiolites) to those from the modern ocean basins (e.g. Salisbury and Christensen 1978).

The Varieties and Flavours of Ophiolites

Ophiolites are diverse in character, with contrasts in their regional geological settings, and more subtle geochemical and isotopic differences. Recent reviews (e.g., Dilek and Furnes 2014; Pearce 2014) define several major categories, and some of these can be subdivided according to their finer details.

The clearest distinction is into so-called 'Tethyan' and 'Cordilleran' types, which have different geological settings. The Mesozoic *Tethyan* ophiolites, including the Troodos complex in Cyprus, and the vast Semail ophiolite in Oman (considered to be the largest known ophiolite; e.g. Searle and Cox 1999; Goodenough et al. 2014) are regionally associated with shallow-water platformal sedimentary sequences. The magmatic stratigraphy of these ophiolites is relatively intact, and they seem to have been transported across passive continental margins. The Paleozoic ophiolites of the Appalachians, although unrelated to the much younger Tethys Ocean, are also of this general type. In contrast, *Cordilleran* ophiolites, exemplified by those of western North America and the wider circum-Pacific region, lack such association with platformal sequences, and are associated with deep-water clastic rocks (turbidites) and low-temperature, high-pressure metamorphism, including blueschists. Their regional tectonic association is with long-lived convergent margins and active subduction zones rather than previously stable continental shelves. Cordilleran ophiolites tend also to be more dismembered and/or deformed, and are typically more altered than their Tethyan counterparts. Ophiolites defined in ancient orogenic belts include examples of both types, but the best-known examples generally correspond to the Tethyan type. Paleoproterozoic ophiolites in northern Quebec (e.g. Scott et al. 1992) likely also belong in this category.

Geochemical data indicate that although there are a few ancient and modern ophiolites that have mid-ocean-ridge basalt (MORB) affinities, most were not formed in such settings. Miyashiro (1972) pointed out that basalts from the Troodos ophiolite in Cyprus had arc-like geochemical traits, suggesting formation in an island arc or back-arc basin, rather than their proposed ridge-type setting (e.g. Gass 1968). This seems to be a general trait, although some larger ophiolites (e.g. Semail in Oman) may record both early ridge-type and later island-arc-type components (e.g. Pearce 2014; Goodenough et al. 2014). The geochemical patterns of ophiolites also define several subtypes (e.g. Dilek and Furnes 2014; Pearce 2014) but this geochemical variation appears to be independent of the more fundamental division into Tethyan and Cordilleran types. Most ophiolites are now considered to have formed in fore-arc or back-arc settings. From the perspective of a field geologist the rocks from these various ophiolite flavours look essentially the same, but the inference of arc-related settings is very important for ideas about the mechanisms of emplacement, as discussed below.

The Emplacement of Ophiolites

Density contrasts between broadly granitic continental crust and mafic oceanic crust indicate that *obduction* of the latter, rather than its subduction, demands special circumstances. Under normal circumstances, dense ocean crust should descend below less dense continental crust, as in subduction zones. However, the attempted subduction of continental crust during terrane accretion or continental collision provides a mechanism by which buoyant continental crust could 'rebound', carrying with it parts of the upper oceanic plate. In simple terms, trying to subduct continental crust beneath oceanic crust is akin to trying to push a buoyant inflatable cushion beneath the water in a swimming pool. This general premise is central to most models for ophiolite emplacement (e.g. Searle and Stevens 1984; Dilek and Furnes 2014). Tethyan-type ophiolites would be emplaced across continental margins when the latter enter a subduction zone that dips *towards* the adjoining ocean, rather than beneath the continent (Fig. 3). Models of this type also explain stratigraphic evolution from a stable shallow-water sequence to deeper-water clastic sedimentation as the volcanic arc approaches and sheds its detritus over the foundering continental shelf (e.g. Church and Stevens 1971; Williams and Stevens 1974; Cawood and Suhr 1992). Intervening sedimentary environments from the outer continental shelf and the continental slope are then assembled tectonically ahead of and below the oceanic crustal slab as it approaches and sheds detritus across them, and are progressively transported across the former continental shelf. The end result is an ordered sequence of tectonic slices in which the distance of tectonic transport increases from bottom to top, culminating in the ophiolite section. The actual assembly and transport of all these components is diachronous, and ophiolite emplacement is the final step. In a general

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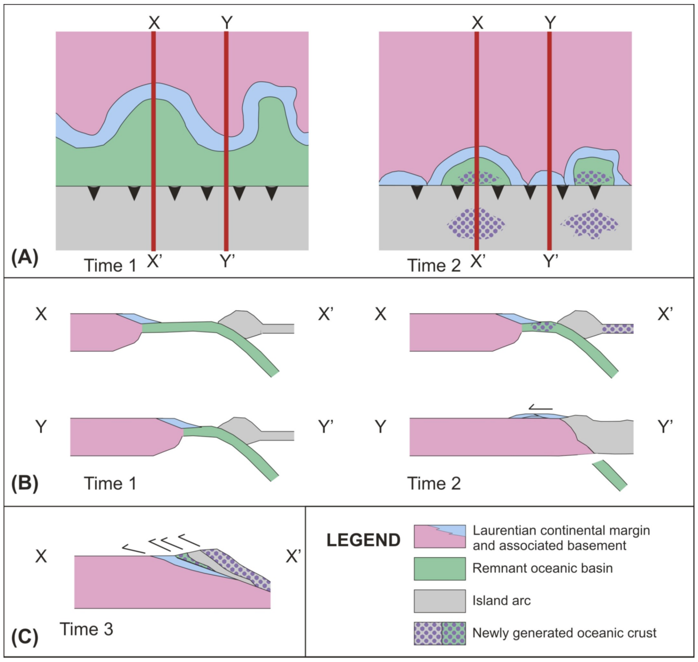

Figure 3. Possible tectonic setting for generation and emplacement of Tethyan-type ophiolites, such as the Bay of Islands Igneous Complex, in which an irregular continental margin approaches a curvilinear subduction zone that dips away from the continent. (A) Ophiolites form in places where oceanic crust remains trapped in embayments (also termed 'reentrants') of the continental margin after adjacent promontories (also termed 'salients') enter the subduction zone. (B) The tectonic architecture along lines X–X' and Y–Y' is shown for Time points 1 and 2. At Time 2, continued subduction of the trapped oceanic crust and fore-arc extension generates new oceanic crust, which is hot and buoyant, and is emplaced onto the continental margin at Time 3 (C), in conjunction with adjoining sedimentary and igneous rock assemblages. For further details and discussion of emplacement mechanisms, see Cawood and Suhr (1992) and Dilek and Furnes (2014). The very schematic diagram is based on models of Cawood (1991) and Cawood and Suhr (1992).

sense, this is the prevailing model for the emplacement of the BOIC and also for the wider geological history of Gros Morne National Park.

Cawood and Suhr (1992) suggested that the generation of the ophiolites and their eventual emplacement both resulted from the irregular geography of the ancient continental margin compared to that of the offshore island arc (Fig. 3). In this model, promontories along the margin (also known as 'salients') will be the first to encounter the subduction zone, but oceanic crust will remain in other areas where there is an embayment in the margin (also known as a 'reentrant'). In such areas, the oceanic crust continues to subduct, leading to its stretching (extension) and perhaps its eventual detachment from the associated continent, promoting the formation of 'new' and hot oceanic crust. It is this newly-formed and relatively buoyant oceanic crust that is eventually obducted onto the approaching continent when remnant basins are finally closed (Fig. 3). In detail, the model is far more complex than these few sentences would suggest, but it is applicable to other ophiolites, such as those of Oman (e.g. Cawood 1991). For further details of other emplacement models and discussions related to other ophiolites, readers should consult Dilek and Furnes (2014) and its extensive bibliography. Although many ideas about ophiolites are now well accepted, there remains much debate about the exact mechanisms of their emplacement.

GEOLOGICAL FRAMEWORK

Gros Morne in the Context of the Appalachian Orogenic Belt

Gros Morne National Park is located on the west coast of Newfoundland, about 70 km north of Corner Brook (Fig. 1). The island of Newfoundland is part of the Appalachian–Caledonian Orogenic Belt, which contains rocks ranging in age from Precambrian to Carboniferous, recording multiple orogenic events of broadly early to middle Paleozoic age, i.e. Ordovician, Silurian, Devonian and Carboniferous. This large orogenic belt was dispersed when the modern Atlantic Ocean opened during the Mesozoic, and other parts of it now occur in northwestern Europe and Scandinavia, the east coast of Greenland, and in various Arctic islands (Fig. 4).

The Appalachian–Caledonian Orogenic Belt was formed when the continent of North America (Laurentia) was amalgamated with present-day Europe and parts of present-day Africa and South America (part of a larger continental assembly called 'Gondwanaland'). This took place in several discrete stages, and the first stage involved the closure of a major ocean basin that existed from the late Precambrian to the late Ordovician. This ocean, which all but disappeared in Silurian times, is now called Iapetus, honouring the father of Atlantis in Greek mythology (Harland and Gayer 1972), but the concept of a precursor ocean to the modern Atlantic was made famous by J. Tuzo Wilson, in his classic paper entitled "*Did the Atlantic close and then reopen ?*" (Wilson 1966). He noted that because the line along which the modern Atlantic opened was not the same as that along which its predecessor had closed, pieces of Europe and Africa were left attached to North America, and pieces of North America were left attached to Europe (Fig. 4). His analysis was, not surprisingly, strongly influenced by the geology of Newfoundland. Harold (Hank) Williams had earlier published a paper that outlined the concept of a "two-sided symmetrical system" for the Newfoundland Appalachians (Williams 1964). Newfoundland has huge importance in the Appalachian Orogen because it preserves both sides of the belt, and also the largest surviving pieces of the Iapetus Ocean. It also provides a vital link between the Appalachians and the Caledonian orogenic belt of Britain and Scandinavia (Fig. 4).

During the closure of the Iapetus Ocean, several island arcs were developed over subduction zones along both the Laurentian and Gondwanaland margins of Iapetus. In the middle Ordovician (~ 470 m.y. ago), the Laurentian continent collided with an island arc or arcs along a subduction zone that dipped away from the continent. Laurentia, and its stable continental shelf, was pushed underneath rocks of oceanic affinity, including deep-water sedimentary rocks, igneous rocks formed in the island arcs, and the deeper oceanic crust with its underlying mantle. These oceanic rocks were then transported tectonically for hundreds of kilometres towards the interior of Laurentia to form what are generally called the *Taconic*

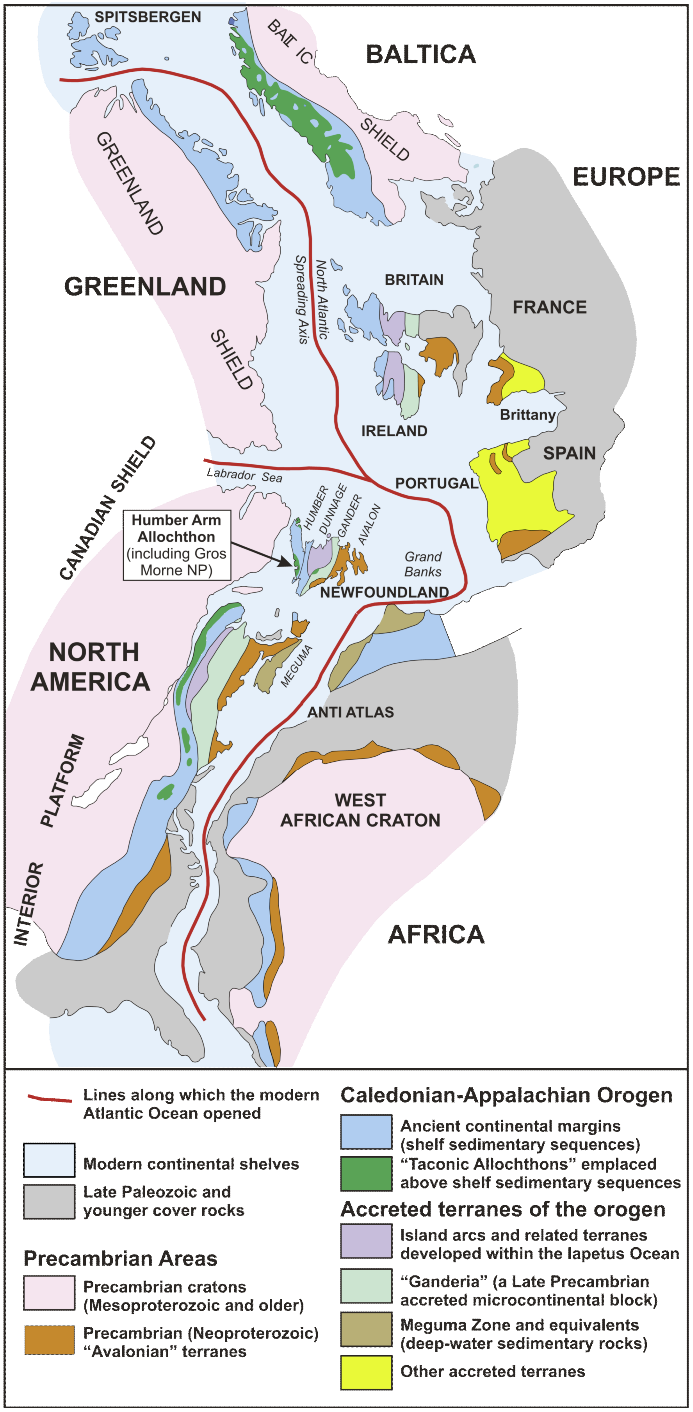

Figure 4. Reconstruction of the Appalachian–Caledonian Orogen and surrounding areas obtained by closing the modern Atlantic and fitting the surrounding continental shelves. Modified from Williams (1995). Note how some parts of modern Europe (e.g. Scotland, Norway) are of North American affinity, and some parts of modern North America (e.g. parts of Newfoundland and Nova Scotia) are of European or African affinity.

Allochthons, named for an area in upstate New York (Fig. 4). The term 'allochthon' (adjective, allochthonous) refers to a collection of rock units that have moved over large lateral distances during and after tectonic collisions. One of the largest of the Taconic Allochthons underlies much of western Newfoundland, including part of Gros Morne National Park. It is known as the *Humber Arm Allochthon* (e.g. Williams and Cawood 1989; Figs. 4 and 5).

Gros Morne National Park is important because it preserves many of these various tectonic elements in one small area. It contains the Precambrian basement rocks of North America and a relatively complete section through the Cambrian–Ordovician continental shelf sedimentary sequence. Sitting structurally above these are the allochthonous rocks transported from the deeper parts of the continental shelf and the Iapetus Ocean, represented by the Humber Arm Allochthon. The Bay of Islands Igneous Complex is the most famous of these transported components. Other allochthonous components include mafic volcanic rocks and plutonic rocks formed in an island arc, and deep-water sedimentary rocks of both clastic and carbonate affinity. The latter represent the ancient continental slope and adjacent basin, and contain remarkable fossil assemblages that help to constrain geologic time (e.g. James et al. 1989). These rocks are in their own way every bit as unique and spectacular as the ophiolites, although they form less prominent landscapes. They include the Global Stratotype Section for the Cambrian–Ordovician boundary, located at Green Point, also within Gros Morne National Park (Cooper et al. 2001).

The evolution of the Appalachian–Caledonian Orogenic Belt did not end with the events recorded in Gros Morne National Park. Volcanism, plutonism and deformation continued through the Ordovician, the Silurian, and well into the Devonian. Gros Morne National Park largely lay outside the influence of these later events, which is another reason for its importance in understanding early events associated with ophiolites. The final chapter in the biography of the Park was written in the Pleistocene, when the spectacular landscape of today was sculpted by glaciation. The most obvious results of glaciation are the much-photographed fjords such as Bonne Bay and Western Brook Pond, but there are lesser features that attest to the power of the ice sheets. Berger et al. (1992) provide some useful summary information on Quaternary history and landscapes, and references to more detailed sources.

An Overview of the Geology of Gros Morne National Park

Gros Morne National Park may be conveniently divided into eight main geological 'packages' (Figs. 6 and 7; after Williams and Cawood 1989; Colman-Sadd et al. 1990; Berger et al. 1992). Two of these packages are in their original locations (i.e. they are *autochthonous*) and the remainder are allochthonous, but some have travelled further than others have. The order of description below corresponds broadly to their structural position (Fig. 7) rather than to their actual stratigraphic age. Subsequent sections provide more background information on the Little Port Complex and the Bay of Islands Igneous Complex (Fig. 5), which are the main foci of the suggested excursions.

Precambrian Basement Rocks (Long Range Inlier)

The oldest autochthonous rocks in the park are Precambrian gneisses and granites of the Long Range Inlier (Fig. 5), which underlie most of the high, barren land in the west (Fig. 6).

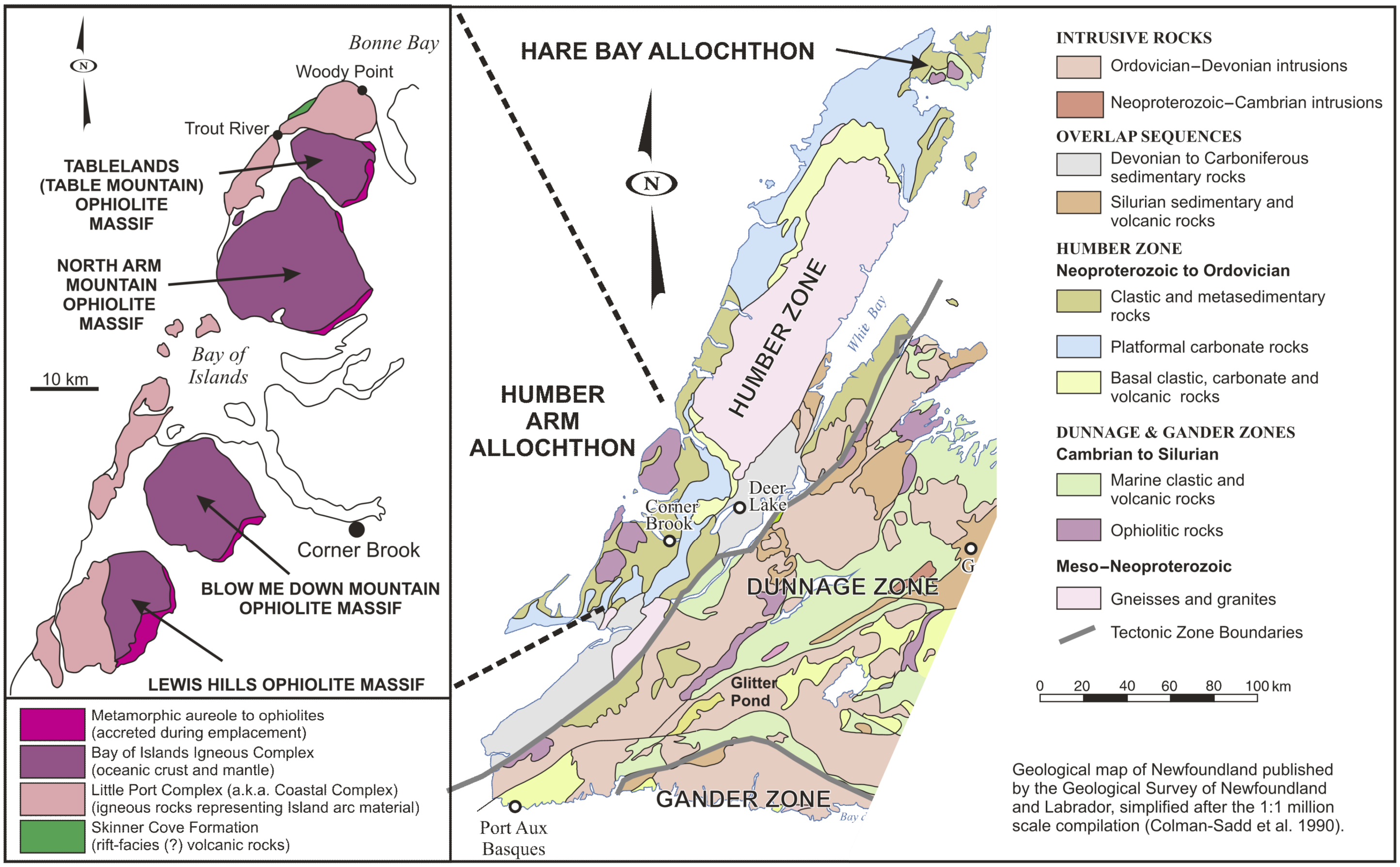

Figure 5. Generalized geology of part of Newfoundland and a more detailed inset view of the ophiolite massifs and related rock units in the Humber Arm Allochthon. Geology after Colman-Sadd et al. (1990), Williams and Cawood (1989), Cawood and Suhr (1992). Note that several other ophiolite sequences exist in central Newfoundland, where they represent parts of the Iapetus Ocean and related basins associated with island arcs.

These rocks were formed well over a billion years ago; they form part of the region of the Canadian Shield known as the Grenville Province. These basement rocks are unconformably overlain by Lower Paleozoic sedimentary rocks in the south of the Park, but are separated from them by west-directed reverse faults in the north (Fig. 7). This west-directed transport and uplift is late (Silurian or Devonian in age) compared to the initial Middle Ordovician emplacement of the allochthons.

Autochthonous Cambrian and Ordovician Sedimentary Rocks

Autochthonous Cambrian and Ordovician strata unconformably overlie the Precambrian basement (Figs. 6 and 7), and record the initiation and development of a stable continental shelf, dominated by carbonate deposition. Gros Morne is not the best place to understand Cambrian and Ordovician stratigraphy, as it has more than a few structural complications, but the succession is relatively complete. The youngest rocks of the autochthonous succession are sandstones that were derived from the east, rather than from continental regions to the west. These youngest sedimentary rocks formed from the erosion of an offshore landmass during early stages of the obduction process (e.g. Church and Stevens 1971; James et al. 1989). This eastward-derived detritus thus records the uplift and erosion of island arcs and ophiolitic rocks during the attempted subduction of the continental margin, as discussed above.

Allochthonous Cambrian and Ordovician Sedimentary Rocks of the Cow Head Group

The Cow Head Group, in the north of the park (Figs. 6 and 7), comprises carbonate strata formed on the outer continental shelf and the continental slope. This remarkable assemblage of deep-water carbonate rocks is correlated with the autochthonous succession using fossils, and it contains the Global Stratotype Section for the Cambrian–Ordovician boundary located at Green Point (Fig. 6; Cooper et al. 2001). The Cow Head Group represents a variety of coexisting depositional environments, but the carbonate rocks in each structural slice are eventually overlain by deep-water clastic rocks derived from the east. The strata of the Cow Head Group thus record the obduction process in a similar manner to the autochthonous sequence (see above), but the change in the sedimentary environment and provenance occurred at an *earlier* time than in the autochthonous rocks, because the allochthonous sedimentary rocks were originally closer to the subduction zone. The inundation of the continental margin by detritus from the east is thus *diachronous*, i.e. it occurred at different times in different locations.

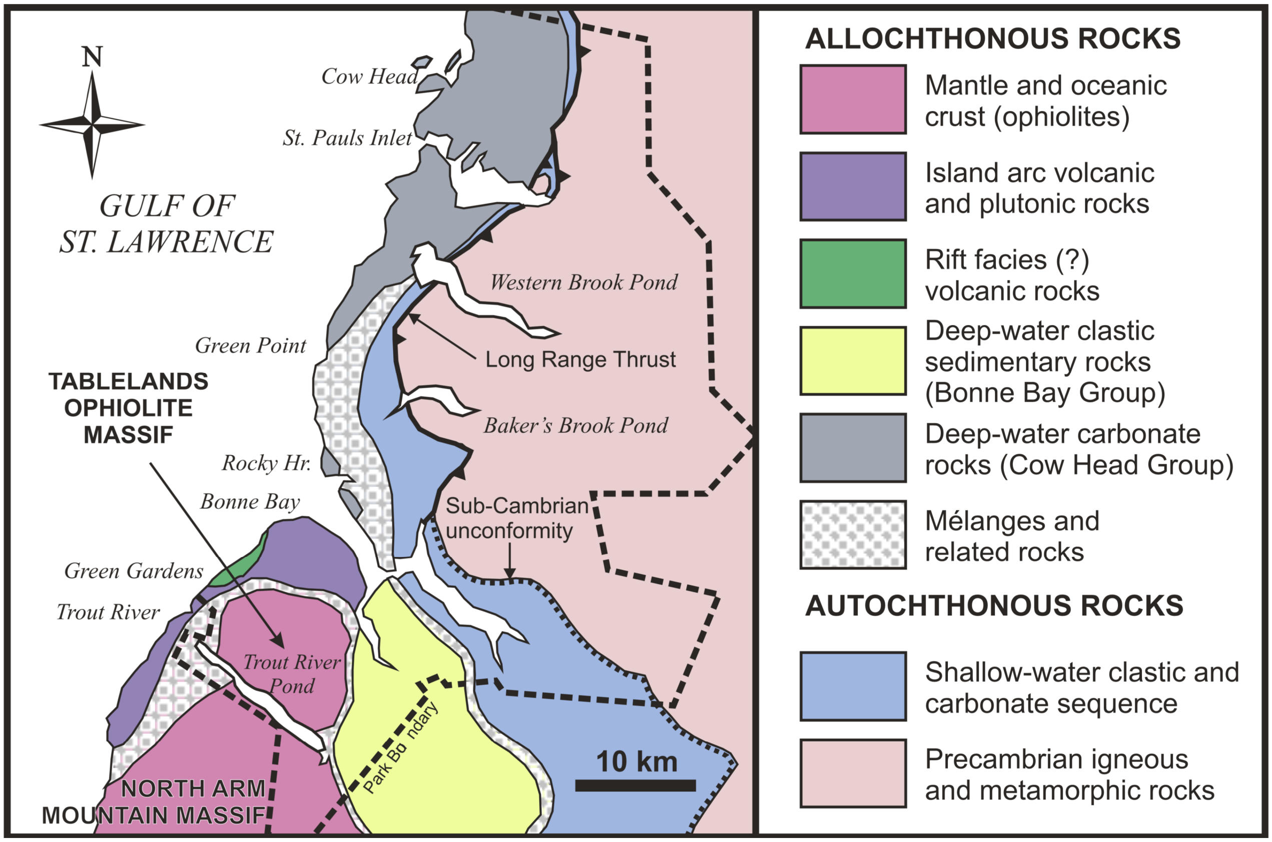

Figure 6. Simplified geology of Gros Morne National Park and surrounding areas, showing major rock assemblages only. Adapted from Berger et al. (1992), Williams and Cawood (1989).

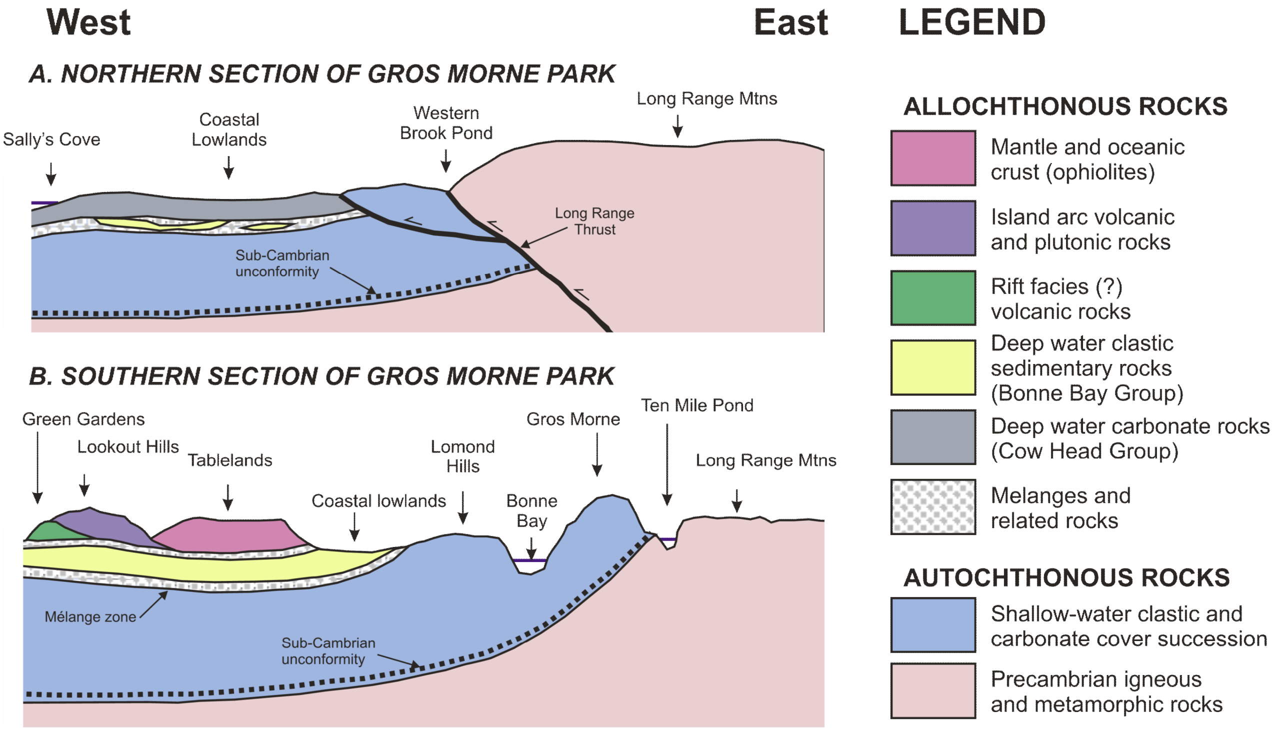

Figure 7. Schematic west to east cross-sections for the northern and southern areas of Gros Morne National Park, showing the inferred structural relationships between autochthonous basement and cover, and various packages of allochthonous rocks, including the Tablelands Ophiolite. Based on Berger et al. (1992), Williams and Cawood (1989).

Allochthonous Cambrian and Ordovician Sedimentary Rocks of the Bonne Bay Group

In the southern area of Gros Morne National Park, allochthonous Cambrian and Ordovician sedimentary rocks (Fig. 6) are mostly of clastic type rather than carbonate-dominated, and are interbedded with lesser mafic volcanic rocks and some deep-water carbonate rocks. These rocks, generally assigned to the Bonne Bay Group, are interpreted to be derived from more distal, deeper parts of the Iapetus Ocean, perhaps underlain by the transition between continental and oceanic crust. Importantly, some sandstones within this package contain detrital chromite, which is a diagnostic indicator for derivation from ultramafic source regions such as the ophiolite suite (Stevens 1965, 1970).

Allochthonous Late Precambrian basalts of the Skinner Cove Formation

Basaltic rocks of the Skinner Cove Formation occur only in a small area of the park, between Bonne Bay and Trout River (Fig. 6) and were originally thought to represent a seamount or oceanic island formed in the Iapetus Ocean, because of their 'alkaline' geochemical traits (Baker 1979). More recent U–Pb geochronology and paleomagnetic data (Cawood et al. 2001; Hodych et al. 2004) indicate a Late Precambrian age (~ 550 Ma). This suggests that they represent older volcanism that might have been associated with the initial rifting of Laurentia and the establishment of a continental margin, as discussed by Cawood et al. (2001).

Allochthonous Igneous Rocks of the Little Port Complex

In the south of the park, allochthonous sedimentary rocks and alkali basalts are in turn structurally overlain by a complex assemblage of igneous rocks, including peridotite, gabbro, diabase, pillow basalt and trondhjemite (Figs. 6 and 7). These rocks are generally termed the Little Port Complex although the more general term 'Coastal Complex' is applied by some (Casey et al. 1985; see also Fig. 5). These are believed to mostly represent higher parts of the oceanic crust developed in an island-arc type setting within the Iapetus Ocean. Minor sedimentary rocks characteristic of deep-water oceanic environments are also present.

Allochthonous Igneous Rocks of the Bay of Islands Complex (Tablelands Ophiolite)

The structurally highest geological package in Gros Morne National Park comprises the spectacular ophiolites of the Tablelands region, representing mostly the upper mantle and the lower sections of the Paleozoic oceanic crust (Figs. 5 to 7). South of Trout River Pond, the ophiolites of the adjacent North Arm Mountain Massif are downfaulted and tilted, and better reveal higher levels of the oceanic crust (gabbro, sheeted dykes and basalts), and also metamorphic rocks at the base of the complex. The metamorphic aureole to the ophiolite is also exposed along the eastern side of the Tablelands Ophiolite, which represents its base. The metamorphic aureoles were probably developed from mélanges consisting largely of mafic volcanic rocks, as initially discussed by Williams and Smyth (1973). The various components of the Bay of Islands Igneous Complex are shown in Figure 5.

Mélange Zones

The remaining geological package in Gros Morne consists of rocks formed *during* the tectonic transport of allochthonous rocks over the autochthonous sedimentary sequences. These generally chaotic and confusing rocks are termed *mélanges* (Figs. 6 and 7). The most extensive belt is in the Rocky Harbour area, and consists of a wide variety of coherent blocks, ranging from a few metres to several hundred metres in size, encased in a shaly matrix. The belt continues on the south side of Bonne Bay. Similar (but thinner) mélange zones separate higher structural slices in the southern part of the park. The mélanges developed under generally cool conditions, but high temperatures and pressures prevailed at the base of the ophiolite sequences, where hot mantle peridotites were exhumed (see previous section).

Geometric Arrangement of Autochthonous and Allocthonous Rocks

The general arrangement of the various rock packages described above is illustrated by schematic cross-sections (Fig. 7). Allochthonous rocks always sit structurally above autochthonous rocks, except in the north of the park, where later reverse faults locally place Precambrian basement above the Cow Head Group. In the region south of Bonne Bay, the stacking order is seen most clearly. The allochthonous sedimentary rocks of the Bonne Bay Group form the base of the stack, with the Little Port Complex and the Tablelands ophiolites sitting structurally above them. The Cow Head Group is absent here, but would sit below the Bonne Bay Group. The exact position of the Skinner Cove Formation relative to the Bonne Bay Group is more difficult to establish, although it clearly sits below the Little Port Complex. The sequence from bottom to top within the allochthonous rocks also represents a geographic sequence, in which higher slices travelled progressively greater distances from the east, across the ancient continental margin represented by the autochthonous sequence.

Ophiolites in Western Newfoundland - A Brief Research History

The history of research in the ophiolites of western Newfoundland is an interesting topic in its own right, and cannot be fully explored here. The ultramafic and mafic rocks were noted by early mappers, but little attention was paid to them until the 1930s, when chromite concentrations attracted more attention (e.g. Snelgrove 1934). Other studies during this time were completed by Ingerson (1935), Cooper (1936) and also by a young geologist from Princeton named Harry Hess. Hess would later be known as one of the founding figures in the plate tectonics revolution, and an interesting paper from that period (Hess 1938) revealed that he thought at length about the wider puzzle of the ultramafic rocks, in addition to the details of local geology (e.g. Buddington and Hess 1937). All early workers considered the rocks to be developed as part of a large layered intrusion or intrusions. Smith (1958) completed the first regional mapping of the entire igneous complex and his synthesis followed the same premise.

Stevens (1965, 1970) provided the first recognition that the Bay of Islands Complex was a well-defined ophiolite suite, and was tectonically emplaced above the sedimentary rocks formerly considered to be older country rocks. There followed a long period of detailed investigation, involving workers from Memorial University and several other institutions, aimed both at understanding magmatic processes in the oceanic crust, and also the later structural and tectonic evolution of the complex. Calon et al. (1988) provide more complete details of all this research, including Ph.D. research projects from that period. Important contributions were also made by Dewey and Casey (2013) and Casey et al. (1985). The work of Suhr (1991) was particularly important in developing models for ophiolite emplacement, as discussed above (e.g. Cawood and Suhr 1992). Compared to the 1980s and 1990s, relatively little research has been completed in the area during the 21st century. Current research activity from Memorial University is centred on alkaline springs, where serpentinization reactions are producing methane (Szponar et al. 2013; Morrill et al. 2014); such environments are considered possible analogues for the surface on Mars.

Allochthonous Igneous Rocks of the Little Port Complex

Most of the localities discussed for the excursions are assigned to the Bay of Islands Igneous Complex (Figs. 5 and 6), but some stops in pillow lavas, sheeted dykes and trondhjemites are actually within the structurally underlying Little Port Complex. The following outline is drawn from Williams and Malpas (1972), Malpas (1979) and Jenner et al. (1991).

The Little Port Complex contains a bewildering assortment of plutonic and volcanic rocks, largely of mafic composition. Amongst the plutonic rocks, the oldest are foliated gabbros, peridotites and their altered equivalents. These were intruded by quartz diorites and trondhjemites. Plutonic rocks of all composition were cut by mafic dykes, of which there are probably several generations. The volcanic rocks of the Little Port Complex formed in submarine environments, and are dominated by pillow lavas and pillow breccias, associated with red cherts and black shales. Minor silicic volcanic rocks are reported in the area south of the park, and hydrothermal alteration of the type associated with sulphide mineralization occurs locally.

The confusing geology of the Little Port Complex is further complicated by alteration and low-grade metamorphism of many of the rocks, such that their original character is hard to establish. Nevertheless, many of its components likely represent the upper layers of the oceanic crust, and they are similar to rocks found locally within the Bay of Islands Complex. U–Pb dating of the trondhjemites near Trout River gave a Cambrian age of ca. 505 Ma, which contrasts with the younger Ordovician age (ca. 484 Ma) of the adjacent Bay of Islands Igneous Complex (Jenner et al. 1991). Based on this, and on aspects of its geochemistry, the Little Port Complex was reinterpreted as part of an older island arc assemblage (Jenner et al. 1991).

Allochthonous Igneous Rocks of the Bay of Islands Igneous Complex (Tablelands Ophiolite)

The Tablelands Ophiolite is the northernmost of four massifs of ultramafic and mafic rocks that form the Bay of Islands Igneous Complex (Fig. 5). A detailed review of all research is beyond the scope of this article, and the following general summary is derived largely from Calon et al. (1988), Cawood and Suhr (1992) and Williams (1995). Information specific to the Tablelands area is contained in Calon et al. (1988) and summarized in Stevens et al. (2003). The general geology of the Tablelands is indicated in Figure 8, redrawn and simplified from Suhr (1991), and also including work by T.J. Calon.

The Bay of Islands Igneous Complex contains all the components of the ophiolite suite. Ultramafic and gabbroic rocks are most extensive, and these are the only areas that are easily accessible. The lower structural contact of the ophiolites, where unmodified by later faulting, is marked by thin discontinuous zones of polydeformed amphibolites and schists. Individual ophiolite massifs within the Bay of Islands Complex, including the Tablelands Ophiolite, are disposed in synclinal structures that expose mantle rocks around their edges and preserve the higher levels of the oceanic crust in their centres. In addition to mafic and ultramafic rocks, there are minor rocks of trondhjemitic composition, which are generally similar to those of the Little Port Complex. Dunning and Krogh (1986) and Jenner et al. (1991) found that the Bay of Islands Complex is significantly younger than the Little Port Complex, at ca. 490–485 Ma. Casey et al. (1985) had previously proposed that these rocks were synchronous, representing mid-ocean ridge and transform fault environments, respectively. Jenner et al. (1991) also presented geochemical and isotopic data suggesting that the ophiolites formed in a supra-subduction zone environment. As discussed above, Cawood and Suhr (1992) proposed a model in which the relatively young oceanic crust now represented by the Bay of Islands Complex was generated within embayments in the Laurentian margin following initial collision of adjacent promontories with an offshore island arc.

The geology of the Tablelands area is described in detail by Calon et al. (1988) and Suhr (1992); the following summary is derived from these sources, and also from reviews by Malpas (1987) and Stevens et al. (2003). The various rock types are best seen in the east of the area, where there are essentially six units. The basal unit is the metamorphic aureole, which is separated from the ultramafic rocks proper by a mylonitic zone developed in the ultramafic rocks (Suhr and Cawood 1993). This basal mylonitic zone is overlain by a thick section of ‘depleted mantle tectonites’, which are mainly harzburgite with lesser amounts of dunite, and dykes and veins of pyroxenite. The mantle tectonites are so named because they contain strong fabrics developed in a high-temperature environment. The adjective *depleted* refers to the fact that these rocks are believed to have lost partial melt during the formation of the overlying oceanic crust.

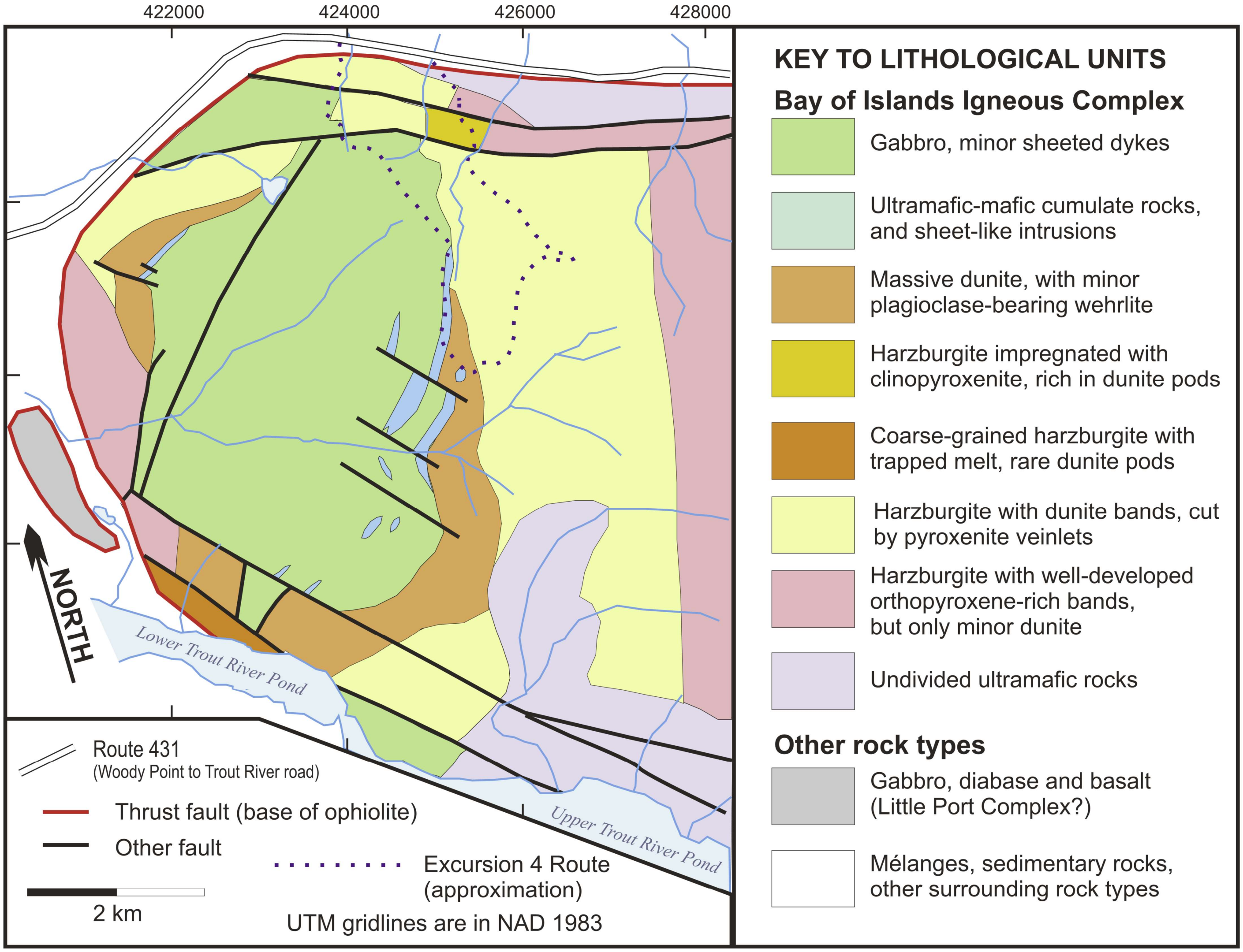

Figure 8. Generalized geology of the western part of the Tablelands Ophiolite in Gros Morne National Park, adapted from a larger map by Suhr (1991), also incorporating work by T.J. Calon included in Stevens et al. (2003). The approximate route of the hike to the Moho area (Excursion 4) is shown, but for navigation, the detailed maps of Figure 9 are required.

The mantle tectonites are overlain by a zone believed to represent originally similar peridotites that were extensively injected by these partial melts. These include harzburgites and abundant dunites, locally with podiform chromite concentrations. This complex unit is sometimes referred to as the 'transition zone'.

The transition zone is overlain by the interesting unit generally termed the 'critical zone', which was so named to reflect comparisons with classic layered mafic intrusions such as the Bushveld Complex and Skaergaard Intrusion (e.g. Smith 1958). This zone consists of strongly deformed ultramafic to mafic cumulates, forming a diffuse boundary between variably melt-depleted mantle rocks and the oceanic crustal rocks that formed from those melts. It represents the Moho, or Mohorovičić Discontinuity. The uppermost unit in the central part of the Tablelands Ophiolite consists of variably deformed to massive or locally layered gabbro, which locally contains diabase dykes.

The Moho is a complex zone, within which the *geophysical Moho* (i.e. the transition from mafic to ultramafic rocks, yielding a seismic velocity contrast) sits a few hundred metres above the *petrological Moho*, which is a more subtle boundary between cumulate rocks and melt-depleted rocks of the mantle. These separate definitions of the Moho are confusing to some (including the author) because both are petrological transitions, but the more obvious lithological contrast is not labelled as such.

The uppermost dyke- and volcanic-dominated parts of the oceanic crust were eroded from the Tablelands, but such rocks do occur within the adjacent North Arm Mountain massif, south of Trout River Pond (Fig. 5). Access to this area without a boat is difficult, as it is rugged and there are no defined trails, so it is not included in suggested excursions. However, the sheeted dykes are clearly visible from viewpoints along the Trout River hiking trail. The Moho is also exposed on the southern shore of Trout River Pond, as is the metamorphic aureole, but these areas are not visible from any of the excursions.

SUMMARY OF FIELD EXCURSIONS

Overview

This article outlines four excursions within Gros Morne National Park, each of which offers a different view of the Tablelands Ophiolite and associated rocks. The most ambitious is Excursion 4, which visits the Moho and underlying upper mantle rocks on the high plateau at around 600 m elevation. This is a strenuous venture and requires special considerations and preparation. Two other hiking excursions explore the lower country around Winterhouse Brook (Excursion 2) and Trout River Pond (Excursion 3), mostly using marked National Park trails. These are part-day ventures that can be completed in less-than-ideal weather, although they are far more pleasant if the sun shines. Excursion 1 is a collection of roadside stops on Route 431 and some associated short optional hikes that allow visitors to see typical rock types, including some that are absent or inaccessible in the Tablelands. Excursion 1, or parts of it, could easily be combined with either Excursion 2 or Excursion 3 to make a full day. However, Excursion 4 requires a full day with a good weather forecast.

Descriptive information for excursions comes in part from previous publications, notably Malpas (1987), Calon et al. (1988) and Stevens et al. (2003). The map of Berger et al. (1992) is an excellent resource with annotated panoramas drawn by the late Ian Brookes, and is highly recommended. However, its large size makes it difficult to use outside on a windy Newfoundland day! Figure 8 provides a geological map for part of the Tablelands Ophiolite and Figure 9 provides topographic maps showing the routes for Excursions 2 and 4. The photographs included in Figures 10 to 32 illustrate most of the outcrops described below.

Safety Considerations and Park Protocols

Visitors to all the excursion areas should be attentive to safety. Roadside outcrops require careful attention to traffic, as Route 431 is narrow and lacks a hard shoulder in some areas. Wherever possible, use the parking areas designated by Parks Canada. Newfoundland's unpredictable weather can be hazardous as well as inconvenient. In high-elevation areas such as the Tablelands plateau, weather can deteriorate rapidly, and fog can obscure routes, leading to disorientation and unexpected encounters with cliffs. Unprepared hikers risk hypothermia if

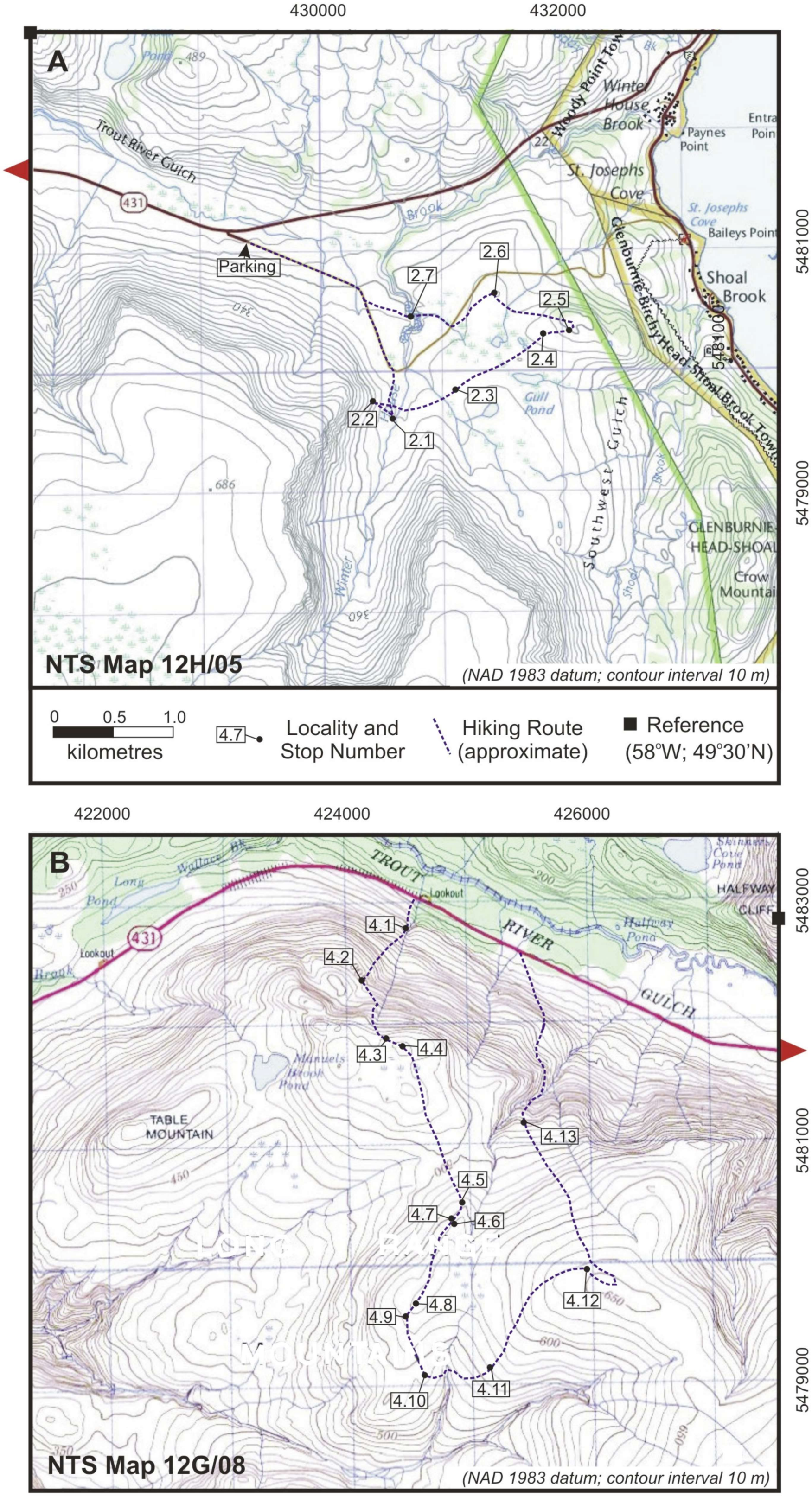

Figure 9. Extracts from National Topographic System (NTS) maps showing the stop localities and hiking routes for Excursions 2 and 4, which in part include off-trail hiking. (A). Part of NTS map area 12H/05 (Lomond) showing Excursion 2. (B) Part of NTS map area 12G/08 (Trout River). The original scale for both maps is 1:50,000, but may differ in this reproduction; the pale blue grid squares represent 1 km on each side. Copyright for the base maps rests with the Government of Canada.

they are caught by changing weather. Even on the marked National Park trails it is wise to have sturdy footwear, with strong ankle support, and this is essential for any off-trail hiking, as the terrain is rugged. In warm sunny summer weather, biting insects will almost inevitably accompany hikers. This article provides descriptive information and maps for off-trail hiking (Fig. 9), but hikers should carry a compass and a general topographic map. The use of a global positioning system (GPS) receiver is strongly recommended, because coordinates are provided for all locations, to allow them to be found precisely. The area is located mostly within National Topographic System (NTS) 1: 50,000-scale map areas 12G/08 (Trout River) and 12H/05 (Lomond), aside from the Woody Point area, which is within NTS area 12H/12 (Gros Morne).

Gros Morne National Park is a protected area, and the collection of samples (including loose material) is prohibited; similar restrictions apply to the removal of plants and flowers, and there are strict regulations about interaction with any wildlife. Formal research activities in the park, even if they do not involve sampling, require approval through Parks Canada. There are facilities for disposal of refuse on most official park trails, but off-trail hikers are responsible for removing any waste to protect the natural environment. Further information on Park regulations and other aspects of visiting can be found on the Parks Canada website (www.pc.gc.ca). Other information connected to visiting the Park is available at www.newfoundlandlabrador.com and its numerous links. Readers are asked to carefully respect all restrictions and procedures that are imposed to protect this World Heritage Site.

Getting to Gros Morne National Park and the Tablelands

Gros Morne National Park is located on the west coast of Newfoundland, north of Corner Brook (Fig. 1). Access is via Route 430, also known as 'the Viking Trail', which departs from Route 1 (the Trans-Canada Highway) at Deer Lake, where the regional airport is located. There is no public transportation to the park so most visitors will need a vehicle; rentals are available at Deer Lake airport and in the City of Corner Brook. Most accommodation and services are on the north side of Bonne Bay, around Rocky Harbour and Norris Point. The drive from these areas to the Tablelands will take slightly more than one hour, depending on traffic. There is also some accommodation on the south side of Bonne Bay, mostly around Woody Point, where the Park interpretation centre is located. Unfortunately, there is no longer a car ferry linking Woody Point and Norris Point, but a summer foot passenger ferry links the two communities, and transportation is also available from Woody Point to the Tablelands trailhead, allowing access to Excursion 3 and possibly Excursion 4 without driving. However, there are only a few return ferry trips each day, and the last water taxi run from Woody Point in the summer of 2019 was at 5.30 pm. Details of the schedule are best obtained onsite in the communities as it may vary from day to day and with the season.

All excursions are on the south side of Bonne Bay, accessed by Route 431, which branches from Route 430 at Wiltondale, and terminates at Trout River. This scenic road initially passes through autochthonous Cambrian–Ordovician carbonate rocks, and then passes into allochthonous transported sedimentary rocks before entering the Bay of Islands Igneous Complex and Little Port Complex between Woody Point and Trout River (see Fig. 6).

Finding the Localities

Roadside outcrops are located using kilometerage from known locations such as junctions or bridges, or by reference to Park facilities, but note that odometer readings in different vehicles may vary by as much as 10%. Stop descriptions include Universal Transverse Mercator (UTM) coordinates derived from GPS readings, which are generally accurate to ± 15 m or so. All UTM coordinates are in Zone 21, and use the NAD 1983 datum. Coordinates can be converted to the earlier NAD 1927 datum by subtracting 63 m from the Easting coordinate and subtracting 219 m from the Northing coordinate. For greater certainty and convenient reference, complete coordinates for NAD 1983, NAD 1927 and latitude/longitude systems are provided in Table 1.

EXCURSION 1: ROADSIDE OUTCROPS AND SHORT HIKES

General Information

This excursion proceeds from east to west, mostly along Route 431, although it does not have to be completed in that direction. Short optional hikes start at the Discovery Centre, at the trailhead for Green Gardens, and near the end of Route 431 in Trout River. Some roadside stops are adjacent to parking areas provided by Parks Canada, but others require careful parking on the gravel shoulder.

Stop 1.1: Panoramic Views of Bonne Bay and the Tablelands Ophiolite (433230E, 5482896N)

From Wiltondale, follow Route 431 to the South Arm of Bonne Bay, passing through the communities of Glenburnie, Birchy Head and Shoal Brook. Stop 1.1 is located just before the junction where Route 431 branches left up a steep hill, signposted for Trout River and the Discovery Centre. It is included because of its spectacular views (Fig. 10). Most geological elements of Gros Morne National Park can be seen from Stop 1.1.

The view of the Tablelands looming over Bonne Bay is justly famous, as is the view of the pyramid-like "Peak of Tenerife", which represents the metamorphic aureole of the ophiolite, sitting beneath and adjacent to the igneous rocks. The mountain was named for its similarity in shape to the high volcano (Mount Teide) that dominates the island of Tenerife in the Canary Islands, off the northwest coast of Africa. "Gibraltar Peak" is the flat-topped mountain across Bonne Bay, and was so named because it resembles the famous Mediterranean fortress. It is made up, in part, of transported volcanic rocks of rift-related origin from the earliest (late Precambrian) structural slices of the Humber Arm Allochthon. These features, and "Table Mountain" itself, were named by Captain James Cook, who was one of the first Europeans to explore the west coast of Newfoundland from 1765 to 1767. The South Arm of Bonne Bay is incised into a regional

Table 1. A compilation of geographic coordinates for all locations discussed in this article, with reference to Universal Transverse Mercator (UTM) systems (NAD1927 and NAD1983 Canada data) and Latitude/Longitude, expressed as decimal degrees and minutes. The coordinates quoted in the article text are all with respect to NAD 1983 datum, but can be converted to NAD 1927 by subtracting 63 m from Easting and 219 m from Northing values. Note that this approximate conversion is valid only for this restricted geographic area.

Stop Identification (see text of paper)	UTM Easting (NAD 1927)	UTM Northing (NAD 1927)	UTM Easting (NAD 1983)	UTM Northing (NAD 1983)	Longitude (degree; min; sec)	Latitude (degree; min; sec)
Excursion 1: Roadside stops and assorted optional hikes						
1.1	433167	5482677	433230	5482896	57; 55; 19.39663 West	49; 28; 13.20201 North
1.2	433837	5483818	433900	5484037	57; 54; 46.78002 West	49; 30; 19.27930 North
1.3	431560	5482815	431623	5483034	57; 56; 39.36309 West	49; 29, 45.90000 North
1.4	432605	5482118	432668	5482337	57; 55; 46.98959 West	49; 29; 23.75213 North
1.5	430810	5481257	430873	5481476	57; 57; 15.66331 West	49; 28; 55.15211 North
1.6	430356	5481157	430419	5481376	57; 57; 38.16110 West	49; 28; 51.72842 North
1.7	422090	5483295	422153	5483514	58; 04; 30.45272 West	49; 29; 57.37263 North
1.8	417798	5481967	417861	5482186	58; 08; 02.79644 West	49; 29; 12.33030 North
1.9	417872	5481578	417935	5481797	58; 07; 58.82839 West	49; 28; 59.77221 North
1.10	418311	5482356	418374	5482575	58; 07; 37.59336 West	49; 29; 25.17524 North
Excursion 2: The upper mantle and aureole around Winterhouse Brook						
2.1	430572	5479490	430635	5479709	57; 57; 26.37465 West	49; 27; 57.84461 North
2.2	430370	5479525	430433	5479744	57; 57; 36.43175 West	49; 27; 58.89493 North
2.3	430955	5479640	431018	5479859	57; 57; 07.44214 West	49; 28; 2.85763 North
2.4	431690	5480100	431753	5480319	57; 56; 31.21336 West	49; 28; 18.04892 North
2.5	432020	5480110	432083	5480329	57; 56; 14.82384 West	49; 28; 18.50538 North
2.6	431385	5480425	431448	5480644	57; 56; 46.56974 West	49; 28; 28.44825 North
2.7	430665	5480265	430728	5480484	57; 57; 22.24324 West	49; 28; 22.97488 North
Excursion 3: Gabbro and peridotite units around Trout River Pond						
3.1	418458	5478951	418521	5479170	58; 07; 27.76096 West	49; 27; 35.00419 North
3.2	420560	5477260	420623	5477479	58; 05; 42.13569 West	49; 26; 41.25596 North
3.3	420860	5477110	420923	5477329	58; 05; 27.13141 West	49; 26; 36.53990 North
3.4	422010	5476365	422073	5476584	58; 04; 29.50011 West	49; 26; 12.95051 North
3.5	422880	5475845	422943	5476064	58; 03; 45.94084 West	49; 25' 56.51053 North
Excursion 4: The lower oceanic crust, moho region, and upper mantle on the high plateau						
4.1	424385	5482570	424448	5482789	58; 02; 35.85827 West	49; 29; 34.92936 North
4.2	424070	5482165	424133	5482384	58; 02; 51.23560 West	49; 29; 21.67655 North
4.3	424255	5481670	424318	5481889	58; 02; 41.69836 West	49; 29; 05.73158 North
4.4	424400	5481595	424463	5481814	58; 02; 34.44025 West	49; 29; 03.36750 North
4.5	424895	5480285	424958	5480504	58; 02; 08.94262 West	49; 28; 21.17102 North
4.6	424830	5480110	424893	5480329	58; 02; 12.05238 West	49; 28; 15.47626 North
4.7	424795	5480160	424858	5480379	58; 02; 13.82561 West	49; 28; 17.07972 North
4.8	424485	5479460	424548	5479679	58; 02; 28.74639 West	49; 27; 54.27814 North
4.9	424425	5479360	424488	5479579	58; 02; 31.65825 West	49; 27; 51.01374 North
4.10	424554	5478860	424617	5479079	58; 02; 24.90628 West	49; 27; 34.88189 North
4.11	425080	5478925	425143	5479144	58; 01; 58.82263 West	49; 27; 37.21857 North
4.12	425900	5479735	425963	5479954	58; 01; 18.63647 West	49; 28; 03.80293 North
4.13	425395	5480925	425458	5481144	58; 01; 44.53440 West	49; 28; 42.11214 North

Note: All UTM coordinates are in Zone 21. Conversion of coordinates between NAD 1927 and NAD 1983 datums, and the calculation of latitude and longitude were completed using the NTv2 online utility program, provided by Natural Resources Canada, at (webapp.geod.nrcan.gc.ca/geod/tools-outils/ntv2.php).
UTM coordinates from GPS measurements are considered to be accurate to +/- 15 m for most localities.

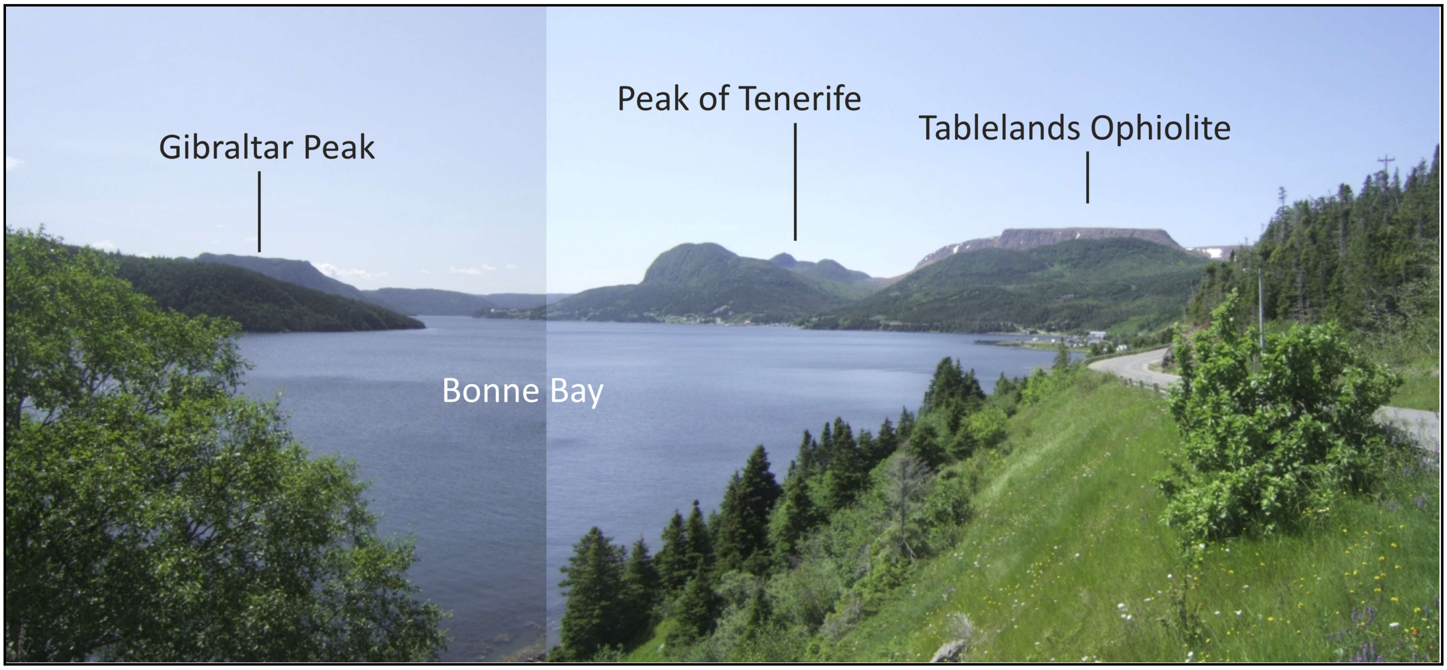

Figure 10. Composite image of the view across the south arm of Bonne Bay, from near Woody Point (Stop 1.1). The view includes part of the Tablelands Ophiolite, the metamorphic aureole beneath it (Peak of Tenerife area) and allochthonous sedimentary rocks of the Humber Arm Allochthon (Gibraltar Peak). The peaks and Table Mountain were named by Captain James Cook after locations he had visited on other voyages.

mélange unit, remnants of which can be seen along its west shore. Gros Morne mountain (elevation 807 m) is clearly visible from the stop, and is the round peak resembling a bald head. Gros Morne is often cited as the highest point on the island of Newfoundland, but is actually the runner-up for that honour; a peak in the Lewis Hills (also part of the Bay of Islands Igneous Complex) is slightly higher at 814 m.

Stop 1.2: Gabbro and Mélange at the Woody Point Lighthouse (433900E, 5484037N)

From Stop 1.1, continue on Route 431 into the village of Woody Point, and take the road nearest to the shore through the small business district. The lighthouse is at the north end of the village, and is easily accessed by a short path. This stop, and several that follow, are located within the Little Port Complex, which sits structurally beneath the ultramafic rocks of the Bay of Islands Igneous Complex.

The Woody Point lighthouse is a much-photographed locality that affords excellent panoramic views of Bonne Bay, including Gros Morne mountain (Fig. 11). It is built upon a relatively homogeneous, fine- to medium-grained gabbro, which is one of the simpler and more homogeneous outcrops in the Little Port Complex. However, the gabbro is actually a large block within the wide mélange zone that sits at the base of the Little Port Complex. The cliffs below the lighthouse expose part of this mélange, which consists of a sheared, broken shale matrix containing lozenges of sandstone. Cracks in some of the sandstone blocks have been injected by the shale matrix material. The cliff exposures are visible from close to the lighthouse, but there is no safe route to descend to the shoreline.

Stop 1.3: Lookout Hills Trail, Tablelands Ophiolite Views, and Pillow Lavas (431623E, 5483034N)

From Stop 1.2, return through Woody Point, and take Route 431 (now signposted for Trout River), which climbs up the hill behind the village, and leads to the Discovery Centre. The Discovery Centre is the main interpretation location in this region of the park, with much of interest to visitors and abundant parking. The Lookout Hills Trail is a short but steep route that leads from the Discovery Centre to a nearby hilltop. The trail provides magnificent views, and some outcrops of variably preserved pillow lavas and pillow breccias. The viewing platform is also an excellent place to eat lunch on a nice day. The 5 km roundtrip trail involves an elevation gain of about 300 m and should only be attempted by visitors in good physical condition; about two hours should be allowed for the roundtrip hike.

The seemingly endless ascent leads to the junction for the summit loop, where the right hand branch leads quickly to the summit viewing platform. The trail junction has exceptional views of the Tablelands, and is a good place to see and photograph the view in the evening light, when the ultramafic mountains may appear blood-red (Fig. 12). The view includes the canyon of Winterhouse Brook, which is part of the hiking route for Excursion 2. At the summit (corresponding to the listed coordinates) the outcrop behind the wooden viewing platform consists of deformed pillow lava, forming part of the Little Port Complex. A short distance further along the trail is a rocky area where several large blocks have been arranged in circles for making fires. Some of these blocks are actually entire pillows, with chilled exteriors and radial cooling cracks.

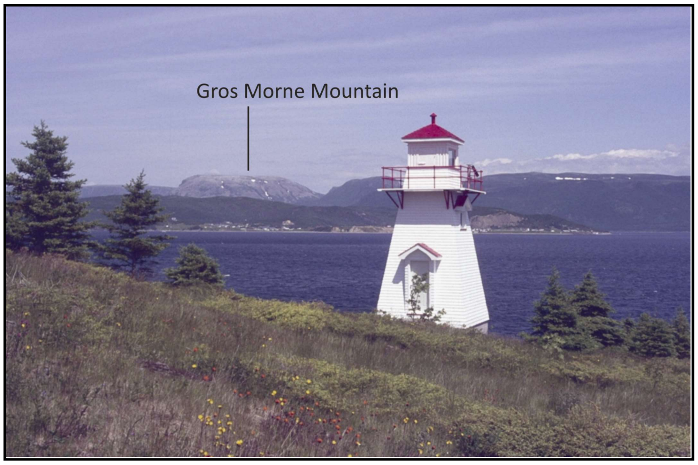

Figure 11. The lighthouse at Woody Point (Stop 1.2) built on a mélange unit including large blocks of altered gabbro (Little Port Complex). Gros Morne mountain (807 m) is seen clearly in the background above the community of Norris Point, across Bonne Bay.

Figure 12. The Tablelands Ophiolite and the deep valley of Winterhouse Brook, as seen from the highest viewpoint on the Lookout Hills Trail (Stop 1.3), above the Gros Morne Park Discovery Centre, near Woody Point. The area around Winterhouse Brook is the location of the Excursion 2 hiking route.

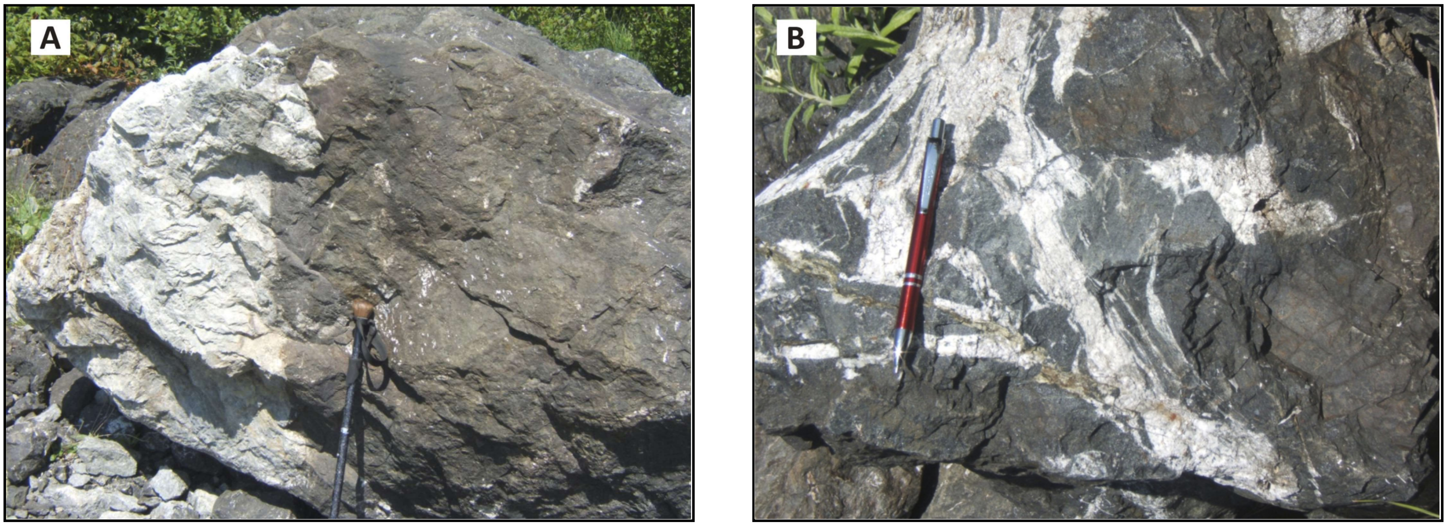

Figure 13. Mafic and granitoid rocks of the Little Port Complex at Stop 1.4, near the Gros Morne Discovery Centre. (A) Vein of trondhjemite (plagioclase-rich granite, at left) cutting altered and weakly metamorphosed gabbro and diabase. (B) Thin trondhjemite veins typically found all through this outcrop.

They have presumably been derived from outcrops by frost action. From the summit area, the trail eventually returns to the trail junction, and then back to the Discovery Centre via a series of boardwalks.

Stop 1.4: Varitextured gabbro, diabase dykes and trondhjemite (432668E, 5482337N)

From the Discovery Centre, walk or drive on Route 431 towards Trout River. This large and very obvious quarry is a very short distance uphill from the Discovery Centre, and was the source of most of the large blocks used in landscaping around the site. A few blocks remain, but the area is now easily accessible. The quarry face should be approached with caution, and many features are better seen in the blocks closer to the road.

The outcrop consists mostly of mafic intrusive rocks of the Little Port Complex (Fig. 13), and these exhibit wide variations in grain size, composition, texture and degree of alteration. Parts of it appear amphibolitic and are weakly foliated. It also contains white-weathering trondhjemite veins, but most of the examples in the outcrop face are thin and exposed on fracture surfaces. In addition to coarse-grained plutonic rocks, the outcrops contain a few areas of mafic dykes, representing higher layers of the oceanic crust, but these are not easy to find. These dyke-like units are mostly fine-grained gabbroic rocks rather than true diabase. This complexity is typical of many outcrops within the Little Port Complex. Although the overall appearance is very messy, this innate heterogeneity adds to the visual interest of the material when it is used for landscaping. Wider grey trondhjemite veins and dykes are most easily observed in large blocks at the uphill end of the quarry (Fig. 13), where there are also blocks of possible volcanic origin. The large blocks are the best places to see intrusive contacts between trondhjemite and the older gabbro, as the out-

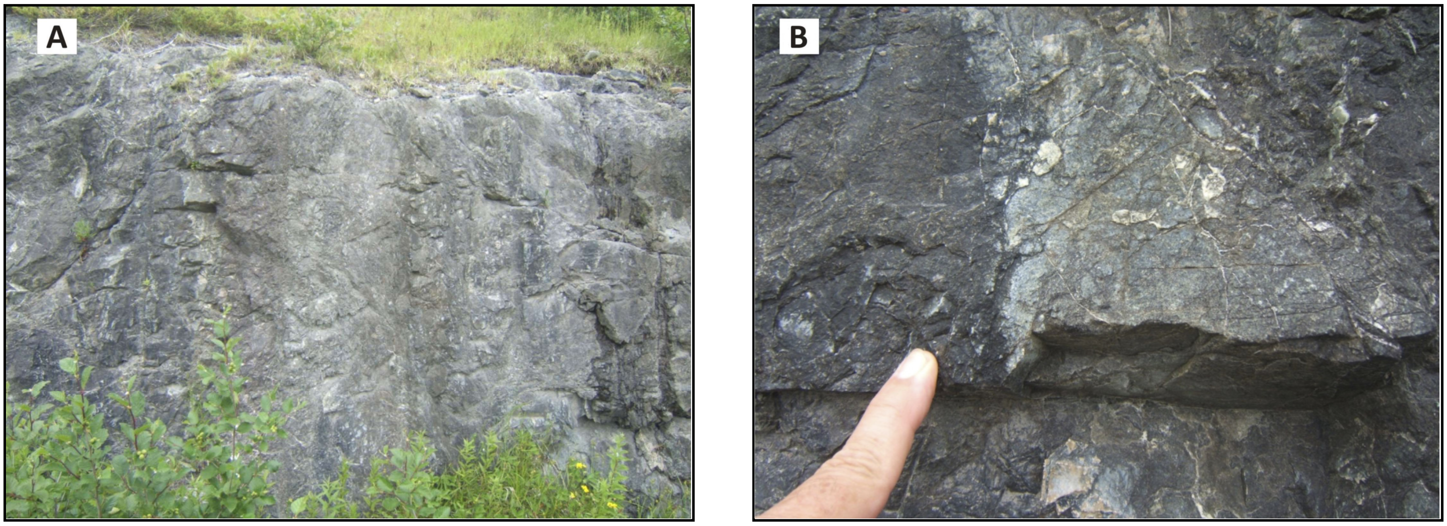

Figure 14. Sheeted diabase dykes found at Stop 1.5, along with possible metavolcanic rocks, all part of the Little Port Complex. (A) General view of outcrops showing colour contrasts. (B) Detail of internal contact zone between two dykes.

crop is very fractured and loose material from the quarry face could be hazardous.

Stop 1.5: Sheeted(?) Diabase Dykes (430873E, 5481476N)

From Stop 1.4, continue on Route 431 for about 1.9 km, to a small outcrop located on the north side of route 431. Park as far off the road as possible, and watch carefully for traffic.

This outcrop is dominated by diabase and provides an interesting contrast with the next stop (1.6), which includes pillow lavas and deep-water sedimentary rocks. The outcrop is massive and well-jointed (blocky) which is more typical of diabase, and it contains some internal contacts, revealed by slight colour and grain size differences (Fig. 14). This suggests that it may consist of sheeted diabase dykes, typical of the upper part of the oceanic crust, located beneath pillow lavas formed on the seafloor. The outcrop is far from spectacular, but as far as I am aware it is the only easily accessible location that represents this important unit.

Stop 1.6: Pillow Lava and Seafloor Sedimentary Rocks, Hydrothermal Alteration and Sulphide Mineralization (430419E, 5481376N)

This outcrop is located opposite the Tablelands viewpoint on Route 431, the road towards Trout River, about 0.8 km west of Stop 1.5. The viewpoint provides safe parking for vehicles, but be attentive to traffic when crossing the road. The reference coordinate is the east end of the outcrop, but the entire outcrop is more than 300 m in length.

This is an extensive outcrop dominated by altered pillow lava, pillow breccia, red chert and red to black shale (Fig. 15). The outcrop is interpreted to represent the uppermost part of the ophiolitic basement to the island arc represented by the Little Port Complex, and part of its overlying sedimentary cover. The prominent rusty zone in the centre of the outcrop provides evidence of hydrothermal alteration and sulphide deposition, and probably represents part of a small submarine hydrothermal system, which vented mineral-rich hot brines onto the ancient seafloor. The red and black sedimentary rocks immediately east of the sulphide-rich zone are cherts interpreted as seafloor sediment, perhaps formed from these exhaled brines.

The eastern part of this outcrop (corresponding to the coordinates listed above) contains well-preserved pillow lavas, which are west-younging based on their geometry and draping relationships. Compared to the previous stop, there are very few dykes, which is consistent with a higher level in the oceanic crust. There are also sections of pillow breccia in this outcrop, with sporadic entire pillows scattered through them (Fig. 15). These are located in the section between the pillow lavas and the sulphide zone, and also dominate the western end of the outcrop. The pillow breccias formed through explosive quenching of pillows, and through physical disaggregation when partly-solidified pillows rolled down submarine slopes.

Stop 1.7: Trondhjemite of the Little Port Complex (422153E, 5483514N)

From Stop 1.6, continue westward on Route 431, passing the trailhead for the Tablelands Hiking Trail (Excursion 2), and a second parking area which is the starting point for hiking up to the summit plateau of the Tablelands (Excursion 4). This parking area also connects to the Green Gardens Trail, but access to the trail is presently closed due to flooding damage. Continue on Route 431 towards its highest point, and park at the main trailhead for the Green Gardens Trail. The Green Gardens Trail is one of the most spectacular routes in the Park, and allows access to well-preserved volcanic rocks of the Skinner Cove Formation, which sit structurally below the Little Port Complex (see earlier description). A short hike of around 1 km along this trail to the first ridge leads to a trondhjemite outcrop, and also provides excellent views across Trout River Gulch to the Tablelands plateau. The vistas will probably make more of a lasting impression than the outcrop.

After leaving the trailhead, there is very little outcrop. The sides of the trail are lined with many orange and brown peri-

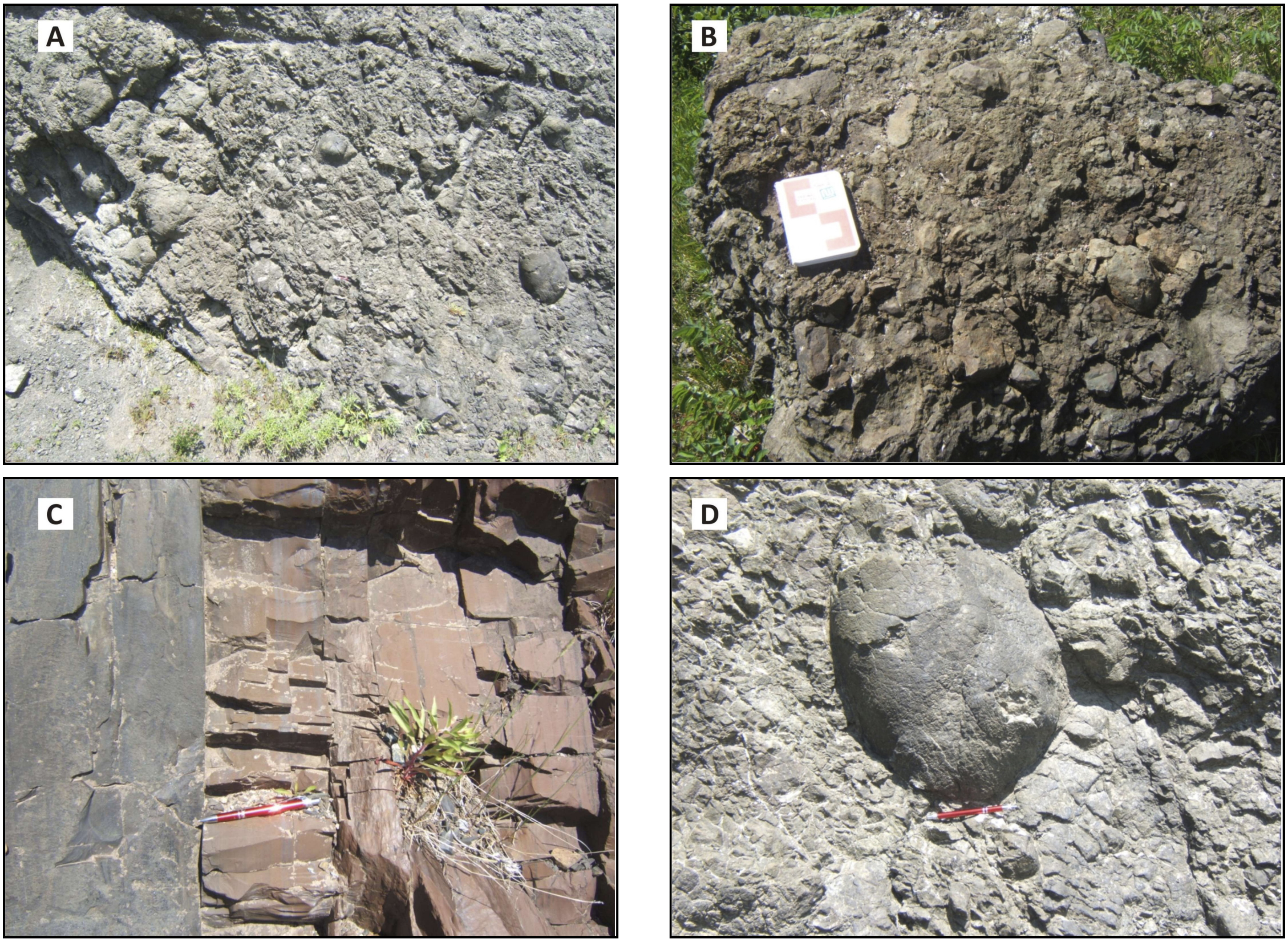

Figure 15. Features of Stop 1.6, an extensive outcrop of pillow lavas, pillow breccias, diabase and sedimentary rocks, all part of the Little Port Complex. (A) General view of pillow lava section. (B) brown-weathering pillow breccias, western part of outcrop. (C) Contact between reddish chert and black chert, both of which are interpreted as seafloor sedimentary facies. (D) A lonely basalt lava pillow within a section of the outcrop dominated by pillow breccias.

dotite boulders and lesser amounts of grey gabbro and white trondhjemite. The landscape is forbidding and barren as the trail climbs gently, and is very typical of the entire Trout River Gulch area (Fig. 16). A few hundred metres from the trailhead, there is a gradual change in vegetation as low bushes and trees appear and cover the hillside. The appearance of this (relatively) lush vegetation probably marks the transition into the largely mafic rocks of the Little Port Complex.

Stop 1.7 consists of obvious white-weathering outcrops on the left hand side of trail, at the first ridge crest. These are coarse-grained, quartz-rich, leucocratic trondhjemite (Fig. 16). The outcrop has the slightly brecciated and fractured appearance typical of much of the Little Port Complex. The mafic minerals have been chloritized, so their original nature is unknown. If you are not continuing on the Green Gardens Trail towards the coast, return to the trailhead by the same route.

Stop 1.8: The Old Man of Trout River (417861E, 5482186N)

From Stop 1.7, continue westward on Route 431 to Trout River, and turn right at the junction on the edge of the village. Continue along the road through Trout River, and turn left on Riverside Drive, across a wooden bridge, and park by the signboards and wooden stairs. A short hiking trail leads from here to a raised seastack known as the "Old Man" (Fig. 17), and also provides access to the Lighthouse Trail (Stop 1.9). The seastack itself is a lineated to foliated amphibolitic gabbro, which includes rusty bands suggesting the presence of sulphides. A smaller collapsed sea stack next to the Old Man is (predictably) known as the Old Woman. The location provides excellent views of glacial features related to sea-level changes (Berger et al. 1992). The following summary is adapted from this source, and also from Brookes and Deardon (1981) and Stevens et al. (2003).

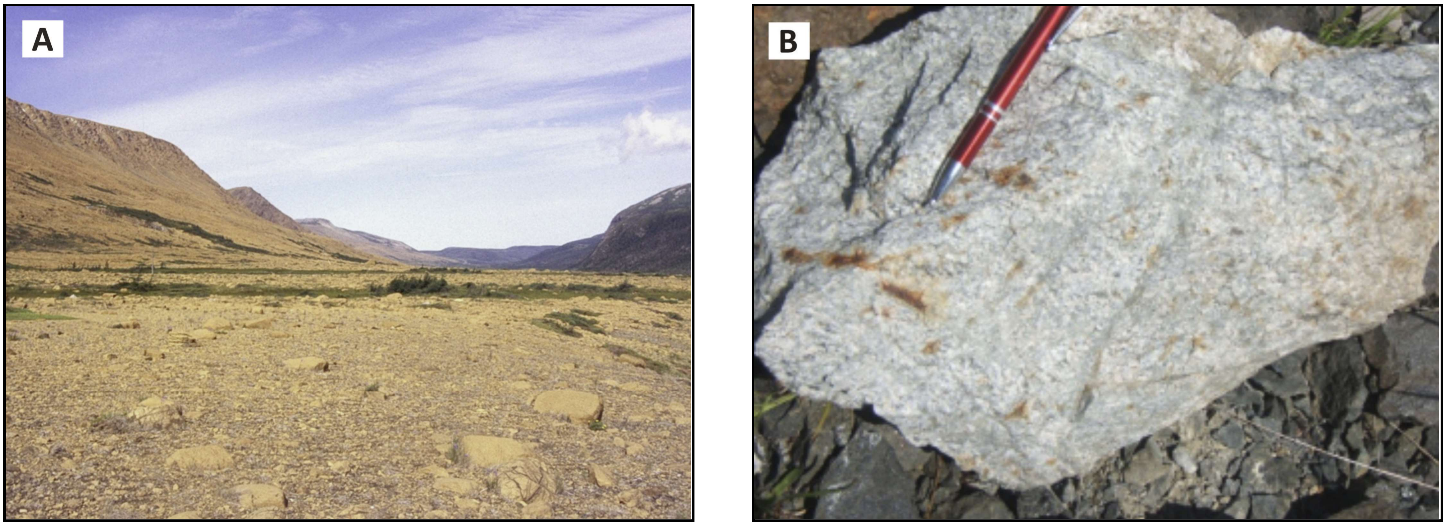

Figure 16. The route through Trout River Gulch. (A) The striking contrast between the barren rocks of the Tablelands Ophiolite (left) and the more 'fertile' wooded hills of the adjacent and structurally lower Little Port Complex (at extreme right). (B) Trondhjemite from the outcrops at Stop 1.7 (an optional short hike) from where there are good views of this area.

Figure 17. The Old Man of Trout River (Stop 1.8), a raised sea stack indicative of a prominent ancient shoreline, seen also is a terrace in the background, upon which most of the village is built.

The village of Trout River is underlain by a large terrace of gravel and sand that now separates the glacially-carved valley of Trout River Pond from the sea. This terrace is a classic ice-marginal marine delta, formed by meltwaters derived from a glacier that existed in Trout River Gulch. The top of this terrace is about 35 m above present day sea level, at the same level as the base of the raised seastack. This former sea level stand at 35 m is also represented by an obvious rock platform that extends on either side of the delta. These prominent features suggest that this was an important and long-lived sea-level stand. Radiocarbon dates obtained from shells indicate that the delta was deposited 12,700 years ago. Some lower terraces in the village mark later stages in the decline of sea level to its present position. Traces of a higher terrace and raised beaches around 70 m above present day sea level are dated at 13,400 years ago.

Stop 1.9: Trail to Trout River Lighthouse (417935E, 5481797N)

Just below the Old Man of Trout River (Stop 1.8) there is a junction on the access trail, signposted for the Lighthouse Trail. This is an easy walk of just over 1 km through some coastal forest and meadows that are full of wildflowers in summer, and leads to an automated lighthouse. There is no outcrop along the trail, but some low cliff-like outcrops visible to the south are part of same raised shoreline at 35 m that is defined by gravel terraces and the raised seastacks. At the lighthouse, the coastal cliffs are visible. These are extensively fractured rocks that are probably of mafic volcanic origin, and are assigned to the Little Port Complex. There are also good views across to the village of Trout River and outcrops on the north side of the harbour, which include some prominent pale grey areas formed from trondhjemites (Fig. 18). These can be visited by hiking along the Eastern Point Trail (Stop 1.10).

Stop 1.10: Trail to Eastern Point (418374E, 5482575N)

There is another short trail from the end of the road at the north end of the beach in Trout River to Eastern Point, which provides access to the trondhjemite outcrops visible from the Trout River lighthouse. To get to the trailhead, return to the main road through the village and turn towards the beach, and then drive or walk to the end of the road. The long wooden stairway ascending the gravel terrace is very obvious at the trailhead, but there is only limited parking. The hike is mostly of interest for its scenery and views, but there are outcrops of lineated coarse-grained trondhjemite at its furthest point (Fig. 18). Other outcrops in the area are black chloritic fractured rocks of uncertain origin, which are superficially similar to those seen in the cliffs at the lighthouse (Stop 1.9).

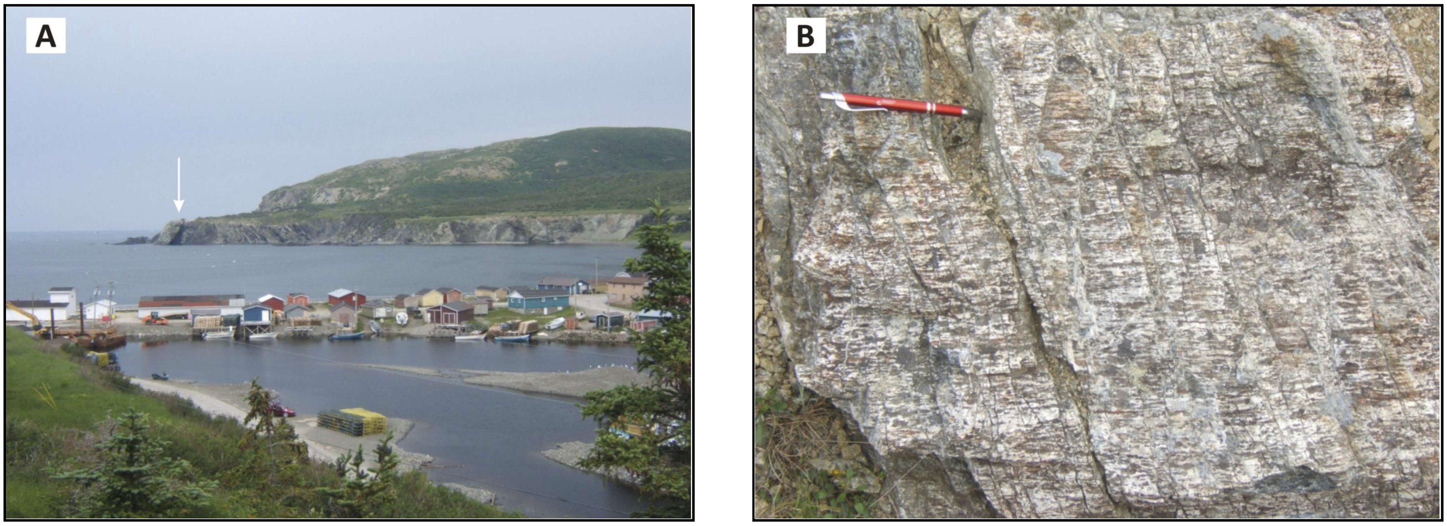

Figure 18. The view from the hiking trail that leads to Stop 1.8 and the Trout River lighthouse (Stop 1.9). (A) The light-coloured bands and areas on the opposite side of the Bay at Eastern Point are trondhjemitic outcrops within the Little Port Complex. Note the prominent raised terrace and shoreline. (B) Strongly lineated trondhjemite from the location on Eastern Point (Stop 1.10) indicated by the white arrow in A.

EXCURSION 2: THE UPPER MANTLE AND METAMORPHIC AUREOLE AROUND WINTERHOUSE BROOK

General Information

A very popular Parks Canada trail commences near Route 431 and leads into the valley of Winterhouse Brook. This provides access to ultramafic rocks that represent the upper mantle section of the ophiolite, and also to much spectacular scenery, locally reminiscent of photos from the Mars Exploration Rovers. Excursion 2 includes this trail and then leaves the trail for some relatively easy off-trail hiking in the area east of Winterhouse Brook, ending at a beautiful river pool. The excursion reveals upper mantle peridotites containing strong tectonic fabrics, variably altered and serpentinized ultramafic rocks, and a small part of the metamorphic aureole that underlies the ophiolite. The Park trail is an easy walk including long boardwalks, but the off-trail hike requires sturdy footwear and adequate clothing, plus a map and (if possible) a GPS receiver. Some of the stops near the Park trail were described in the earlier article by Malpas (1987), but additional information is provided here.

Tablelands Trailhead to Winterhouse Brook

The parking area for the Tablelands interpretation trail is reached via a short branch road from Route 431, about 3 km west of the Discovery Centre. The hiking trail follows an old road for about 1.6 km, and is easy to walk. The route lies very close to the fault that separates ultramafic rocks to the south from mélanges and sedimentary rocks to the north. There are no outcrops along the trail, but there are many large blocks of brown-weathering peridotite derived from the huge cliffs that loom above the valley. Some blocks show spectacular fracture coatings of serpentine minerals. There are excellent views to the west, where the barren landscape of the Tablelands contrasts with the wooded hills of the Little Port Complex on the other side of Trout River Gulch. The flat landscape also shows frost polygons (in which a central region of fine-grained gravelly material is surrounded by a ring of larger angular blocks on a scale of tens of metres). These are not always easy to see from ground level, but are widely developed. They result from frost-heaving and suggest the former presence of permafrost in this high-elevation area; similar frost polygons are widely present in the Canadian Arctic. At Winterhouse Brook, the Park trail diverges south from the old road, and after a few hundred metres it becomes a long boardwalk that leads to a small viewing platform. This is often a busy location in the summer season.

Stop 2.1: Winterhouse Brook Canyon and Alkaline Springs (430635E, 5479709N)

Stop 2.1 is located at the viewing platform at the end of the boardwalk, where there is abundant outcrop of massive brown-weathering peridotite in the stream and around the platform (Fig. 19). These are the most easily accessible peridotite outcrops in Gros Morne National Park, but they are generally homogeneous and featureless compared to the outcrops at Stop 2.2. Thin cross-cutting veins of serpentine minerals are prominent throughout the outcrops and locally yield photogenic fracture coatings (Fig. 19). The site has excellent views up the canyon of Winterhouse Brook, which is the largest of the river valleys that drain the Tablelands Ophiolite. The river course consists of numerous blocks, largely of ultramafic composition, and the bottom and sides of the valley are almost devoid of vegetation. The valley sides consist of unstable scree slopes that can be dangerous and difficult to negotiate. They are best avoided. The brook can be followed up into the canyon for around 1 km, although there is no defined trail, and the numerous peridotite blocks do not make for easy walking. Walking into the canyon is for the most part a scenic excursion, as the nearby outcrops at Stop 2.2 are probably more interesting for geologists.

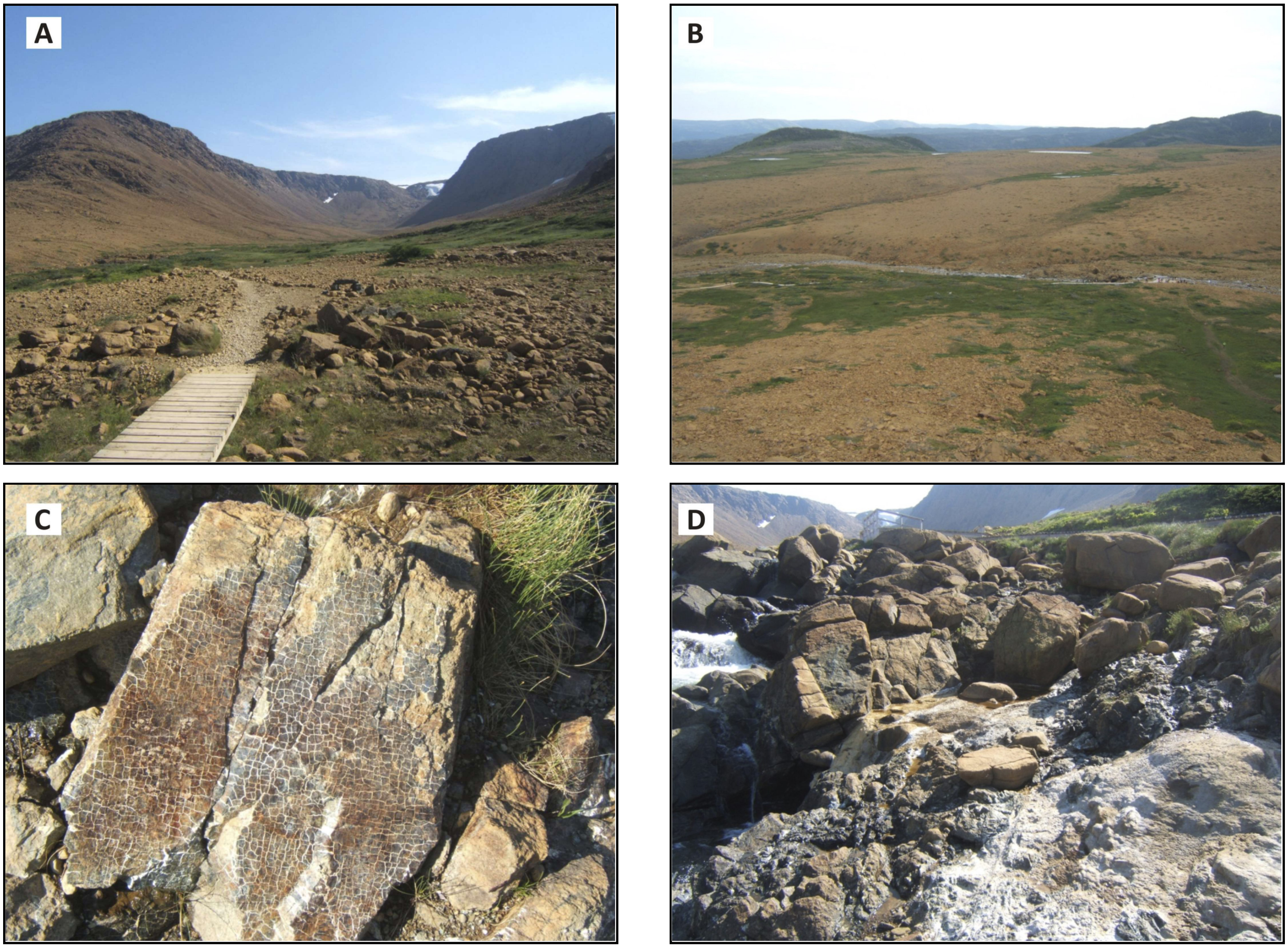

Figure 19. Interesting features of the area around Winterhouse Brook Canyon (Stop 2.1 and vicinity). (A) A view into the canyon from the access trail. (B) View of the desolate landscape around the watercourse, underlain by ultramafic rocks of the mantle, viewing platform at the right hand side, as seen from Stop 2.2 (see Fig. 20). (C) Serpentine minerals on a fracture surface of a peridotite block. (D) Alkaline spring adjacent to Winterhouse Brook, showing white and cream travertine deposits from high-pH waters formed through subsurface serpentinization reactions.

About 75 m north from the viewing platform, the most accessible of several alkaline springs in the Tablelands Ophiolite is located adjacent to the boardwalk on the west bank of Winterhouse Brook. The presence of such alkaline springs was first noted by Stevens (1988). The alkaline groundwaters have deposited white and cream travertine-like layers over peridotite boulders and bedrock at this locality (Fig. 19). These unusual high-pH ground waters are produced by several chemical reactions involved in the formation of serpentine minerals by waters percolating through the ultramafic rocks. Sites such as this are suggested to be possible analogue sites for the surface of Mars, and a possible abiogenic source of methane (e.g. Szponar et al. 2013; Morrill et al. 2014). The springs at this locality are an active research project, and visitors should avoid walking on the travertine surfaces, which are fragile. Note that, although this is a spring, it is NOT suitable as drinking water, as its pH is in the same range as many domestic cleaning agents.

Stop 2.2: Depleted, Strongly Deformed Harzburgites and Pyroxenites, or "Mantle Tectonites" (430433E, 5479744N)

From the viewing platform at Stop 2.1, a faint trail leads west for about 200 m towards the base of the cliffs, providing abundant outcrops, which correspond to Stop 1b in the earlier article by Malpas (1987). The northern edge of these outcrops at the base of the cliff (right hand side as you approach them) is the primary area of interest. The layering in the outcrops becomes increasingly obvious as you hike up from the viewing platform. Layered rocks of similar type continue up the steep hill above the site, but this area can be hazardous and contains much loose rock.

The outcrops (Fig. 20) are the most accessible examples of harzburgites from the mantle tectonite sequence, and they have been strongly deformed under the extreme pressure and temperature conditions of the mantle. The layering is not of magmatic origin, but instead results largely from deformation

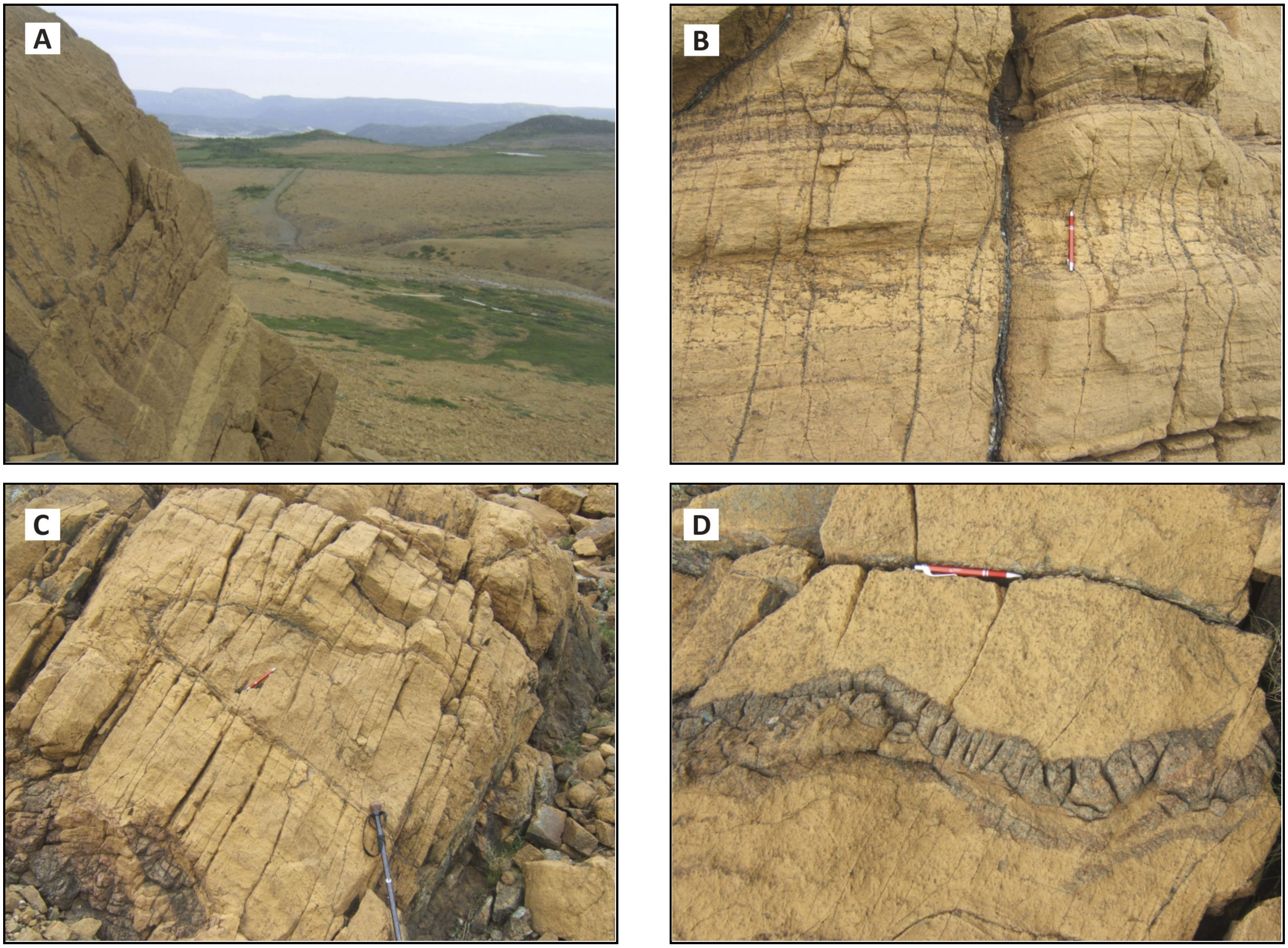

Figure 20. The 'mantle tectonite' outcrops above Winterhouse Brook (Stop 2.2). (A) General view towards Bonne Bay from above the site showing the landscape and the banding in the rocks. (B). Strongly banded harzburgite and dunite with concordant pyroxenite bands (best seen at upper left); the vertical vein consists of serpentine minerals and cuts the older banding. (C) One of several tight to isoclinal folds defined by pyroxenitic layers in these outcrops. (D) Thicker pyroxenite veins showing evidence of some later gentle folding.

processes. The harzburgites contain about 60% olivine and 30% Mg-rich orthopyroxene, with smaller amounts of clinopyroxene and oxide minerals. These are believed to be residual ultramafic rocks that have been depleted in their low-melting point constituents by partial melting related to the formation of the rocks of the overlying oceanic crust. The banded appearance of the outcrops largely reflects variations in the olivine to orthopyroxene ratio, which leads to differences in colour and weathering response. The contacts of individual bands appear abrupt, but they are generally gradational in detail; individual bands can be traced for up to 500 m from this stop. Some of the layers are up to 100 m thick, but most are just a few metres thick. The layering results from intense ductile deformation of original compositional heterogeneities in the rock, and the rotation of veins and dykes of different composition to their hosts. The process that produced the intense layering is called transposition, and the formation of these rocks is closely analogous to the development of banded gneisses in high-grade metamorphic environments. The features evident in these spectacular outcrops are similar to those described from other ophiolite complexes around the world.

The banding in the ultramafic rocks is cut locally by less-deformed veins and dykes of variable composition; some are essentially pyroxenitic, whereas others are extremely olivine-rich and are almost dunitic. The pyroxenite veins and pods are best observed in the higher parts of the outcrop. Pyroxenites are readily identified because they lack the yellow-brown weathering crust developed on the olivine-rich rocks and they exhibit many crystal faces from the pyroxenes within them, which weather to a lustrous pale grey-green. On a sunny day, these pyroxenite veins are easy to see because they actually sparkle. The pyroxenite zones locally retain discordant contacts, but elsewhere are parallel with the regional layering because they have been reoriented by deformation. Some of

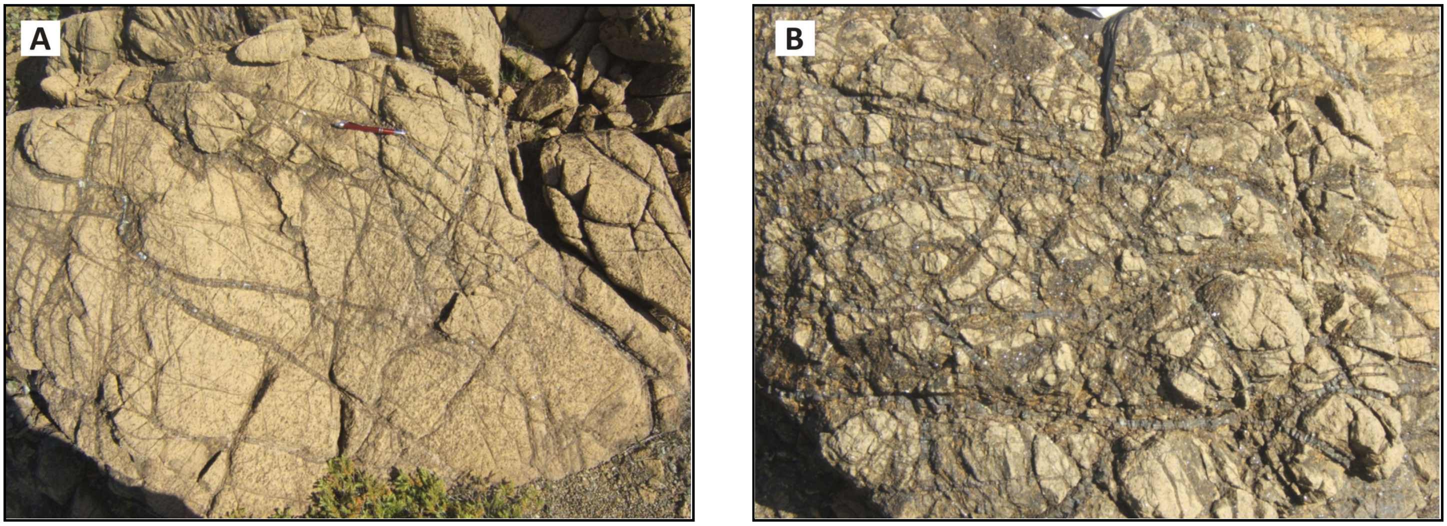

Figure 21. Typical appearance of ultramafic rocks representing the mantle around Winterhouse Brook. (A) Networks of serpentine veinlets dissecting peridotie. (B) More intense development of fractures and serpentine veins. Photos are from Stops 2.3 and 2.4, respectively.

these pyroxenitic bands are folded to form tight to isoclinal folds (Fig. 20), some of which are entirely contained within the regional layering, i.e. they are rootless intrafolial folds.

Both types of veins likely formed during partial melting of the mantle to produce mafic liquids. The dunitic veins have retained only olivine, which was the first mineral to crystallize from these liquids; pyroxenites result from coprecipitation of olivine, orthopyroxene and clinopyroxene. The gabbros in the western section of the Tablelands (see Excursions 3 and 4) represent the final crystallization products of the mafic liquids generated at this deeper level within the mantle.

Stop 2.3: Harzburgites With Serpentine Veining (431018E, 5479860N)

From Stop 2.2, walk downhill again towards the boardwalk, and continue across Winterhouse Brook. The brook is easily crossed at most times of the year. From the brook, walk eastward towards a partially tree-covered hill on the horizon that has a grey colour unlike the yellow-brown that dominates most of the landscape immediately ahead. The terrain is relatively flat and varies between gravelly material that is relatively pleasant to walk upon and areas of angular peridotite blocks that are very hard on the feet and ankles.

The outcrops of interest are located on a low broad ridge, about 0.5 km from Winterhouse Brook. They are harzburgites akin to those of Stop 2.2, but they lack the well-developed compositional layering. These outcrops instead show visually striking net-like patterns of serpentinized veinlets that are very typical of many ultramafic rocks in the Tablelands, and elsewhere (Fig. 21). The pattern is in many respects reminiscent of the veined and dissected appearance of olivine in thin sections of ultramafic rocks, and is an interesting example of how outcrop-scale and microscopic textures can mirror one another in these olivine-dominated rocks.

Stop 2.4: Strongly Serpentinized Ultramafic Rocks (431753E, 5480319N)

From Stop 2.3, continue walking for about 700 m to the prominent low grey-coloured hill noted above. The colour contrast might suggest that it consists of a different rock type to Stop 2.3, but this is not the case. The outcrop (Fig. 21) is more intensely veined and pervasively serpentinized compared to Stop 2.3, which is in turn more serpentinized than the well-preserved rocks at Stop 2.2. This is probably because the excursion route is getting progressively closer to the faulted contact between the Tablelands Ophiolite and adjacent mélange units.

Stop 2.5: Amphibolites of the Metamorphic Aureole (432083E, 5480329N)

From Stop 2.4, walk southeast, crossing some more outcrops of strongly serpentinized ultramafic rocks, and then turn northwards to ascend a partially barren ridge with some small scrubby trees on its upper part. Head towards two distinctive pyramid-shaped peridotite boulders, and then climb as far as you can without actually getting into the trees, which is not advised, as they consist mostly of impenetrable wind-stunted spruce. The upper part of this small hill exposes amphibolites of the metamorphic aureole beneath the ophiolite. Stop 2.5 also provides some wonderful views of the east flank of Table Mountain, and reveals the steep forested mountains above Birchy Head, including the distinctive Peak of Tenerife (Fig. 22). The peak is also part of the metamorphic aureole, although it is not physically connected to the rocks exposed at Stop 2.5. The prominent mountain above Birchy Head consists of altered basalts and dykes representing a large tectonic inclusion of Little Port Complex rocks within the mélange zone. There is also a very good view northward to the South Arm of Bonne Bay, Woody Point and Norris Point.

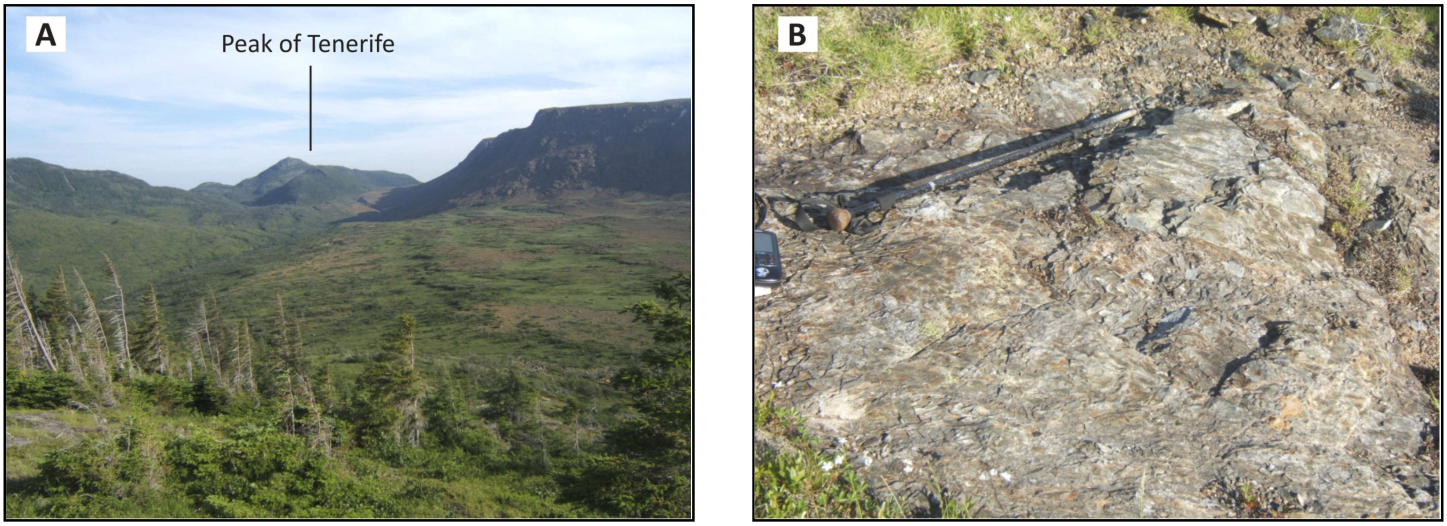

Figure 22. Stop 2.5 located on the metamorphic aureole of the Tablelands Ophiolite. (A) The contrast between the eastern edge of the ophiolite (at right) and the forested hills of the metamorphic aureole, including the Peak of Tenerife. (B) Foliated and lineated fine-grained amphibolitic rock of the aureole, likely derived from mélange including mafic volcanic material.

The rocks are not easily examined, but this is the only portion of the metamorphic aureole that is easily accessible on foot. The rock type is a fine-grained, strongly foliated to lineated amphibolite of mafic composition that locally shows some poorly developed compositional layering (Fig. 22). The original nature of the aureole rocks is not always clear, but they are generally of basaltic composition and are likely derived from rocks of the upper oceanic crust (i.e. sheeted diabase dykes and basalts) in mélanges that were "welded" to the ultramafic rocks during the earliest stages of obduction.

Stop 2.6: Severely Altered and Sheared Ultramafic Rocks (431448E, 5480643N)

From Stop 2.5, retrace the route across Stop 2.4, and then head northwest towards Stop 2.6, which is marked by a low rubbly hill with a prominent green meadow on its west side in the summer months. This spot is mostly included for its superb views of Bonne Bay, including Woody Point, Norris Point and Neddy Harbour, with Gros Morne mountain in the background. The outcrop, which is less spectacular, consists of severely altered and sheared ultramafic rocks, which superficially resemble shale. The outcrop is located very close to the faulted northern contact of the ophiolite, which probably explains why it is strongly sheared and retrogressed.

Stop 2.7: 'Xonotlite Layer' at the Northern Contact of the Ophiolite (430728E, 5480484N)

From Stop 2.6, walk a short distance northwest to the old road, which is marked by an abandoned telephone line. Continue northwestward across the rubbly peridotite and gravel. Just before a band of scrubby vegetation, there is a small stream, flowing westward. Follow this stream to the west. The valley gradually becomes more deeply incised, and eventually leads into the deeper lower valley of Winterhouse Brook, which is canyon-like. The descent along the brook is not difficult, but be wary of large blocks of ultramafic rock that might be unstable. The valley runs just south of the fault that marks the northern edge of the Tablelands Ophiolite, and the outcrops are all strongly serpentinized ultramafic rocks. Some fracture surfaces exhibit spectacular fibrous green and blue-green serpentine minerals (Fig. 23). The stream eventually joins Winterhouse Brook. Stop 2.7 is visible downstream from the junction of the tributary brook with Winterhouse Brook, and it is marked by a prominent wall-like feature extending across the brook (Fig. 23). It corresponds to Stop 1a of Malpas (1987) and marks the contact of the ophiolite with the sedimentary rocks to the north.

The contact of igneous and sedimentary rocks at this location is a fault, and there is no metamorphic aureole preserved. However, boulders of aureole rocks that resemble those of Stop 2.5 are located here and there in the stream bed. The ultramafic rocks in this area are pervasively altered, and few original textures or minerals are preserved. The contact itself is marked by a resistant, pale grey unit that resembles a bed (Fig. 23). This is commonly referred to as the "xonotlite layer", after a rare mineral that it contains. It is actually a metasomatized ultramafic rock that also contains Ca-rich prehnite, calcite and wollastonite. Because the ultramafic rocks contain very little calcium, it is generally assumed that the source of the extra calcium in this layer was *outside* the ophiolite, i.e. within the adjoining sedimentary rocks. The metasomatism of the ultramafic rocks at the contact may have occurred during emplacement of the ophiolite, but could equally well reflect fluid migrations associated with later faulting.

Xonotlite is an unusual mineral with the formula $Ca_6Si_6O_{17}(OH)_2$. It is relatively hard (6.5 in the Moh's hardness scale) which accounts for the resistant nature of this layer. It is named for a place called Tetela de Xonotla, in the state of Pueblo, Mexico. The Winterhouse Brook locality was the first xonotlite to be described in Canada. The xonotlite was initially reported here by Smith (1958), and he provides the most detailed description that I have found. The mineral occurs at

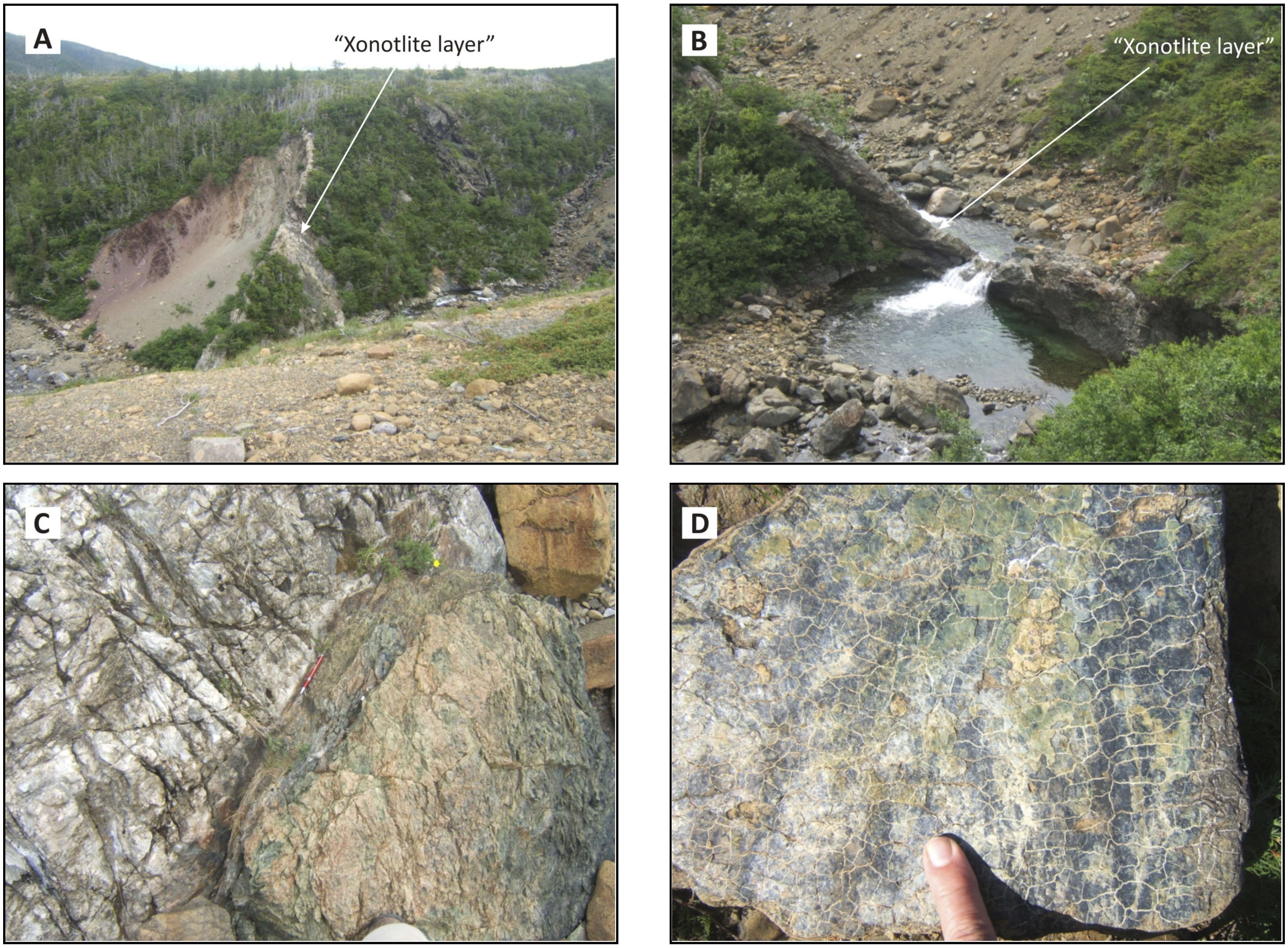

Figure 23. The edge of the Tablelands Ophiolite in Winterhouse Brook, at the 'xonotlite layer' (Stop 2.7). (A) Contact between reddish shale and mélange (left) and serpentinized ultramafic rocks (right), including the resistant xonotlite layer. (B) A view of the xonotlite layer crossing the brook, where it creates a waterfall and a tempting pool; note that access to the pool for swimming is now difficult and potentially hazardous. (C) Contact between the xonotlite layer (white, metasomatized ultramafic rock) and soft green serpentinite at right. (D) One of many spectacular fracture surfaces displaying serpentine minerals seen in this area.

another locality in the Bay of Islands Igneous Complex (Cox's Cove; Smith 1958) and also in calc-silicate metamorphic rocks near Rose Blanche, in southwestern Newfoundland (Brown 1978). Visitors should not expect to collect any samples, because Park regulations prohibit any removal of material. Also, the mineral is essentially invisible to the naked eye, as it is intergrown with the other minerals noted above.

The xonotlite layer forms a prominent waterfall in Winterhouse Brook, below which is a beautiful deep circular pool. Unfortunately, access to this pool for swimming is increasingly hazardous because the banks have become deeply eroded and very steep. Although it may be tempting, it is not recommended.

From Stop 2.7, cross Winterhouse Brook below or above the swimming hole, and carefully climb the western slope of the valley, sticking to the more stable vegetated slopes. At the top of the bank, turn upstream and walk south or southwest to rejoin the old road between Woody Point and Trout River. This section of the hike crosses some very well-developed frost polygons before joining the old road. The walk back to the parking area should take 15 to 20 minutes.

EXCURSION 3: PERIDOTITE AND GABBRO IN THE TROUT RIVER POND AREA

General Information

The Trout River area has several features of geological interest, including the raised sea-stack of The "Old Man" and other indications of changing sea level (see Excursion 1). The long lake of Trout River Pond occupies the valley between the Tablelands and North Arm Mountain, both of which are part of the Bay of Islands Igneous Complex (Fig. 5). Trout River Pond has excellent shoreline exposures that represent the upper crustal section of the latter, and also the Moho region, but these are only accessible by boat. A tourist excursion on the lake during the summer provides a chance to see these

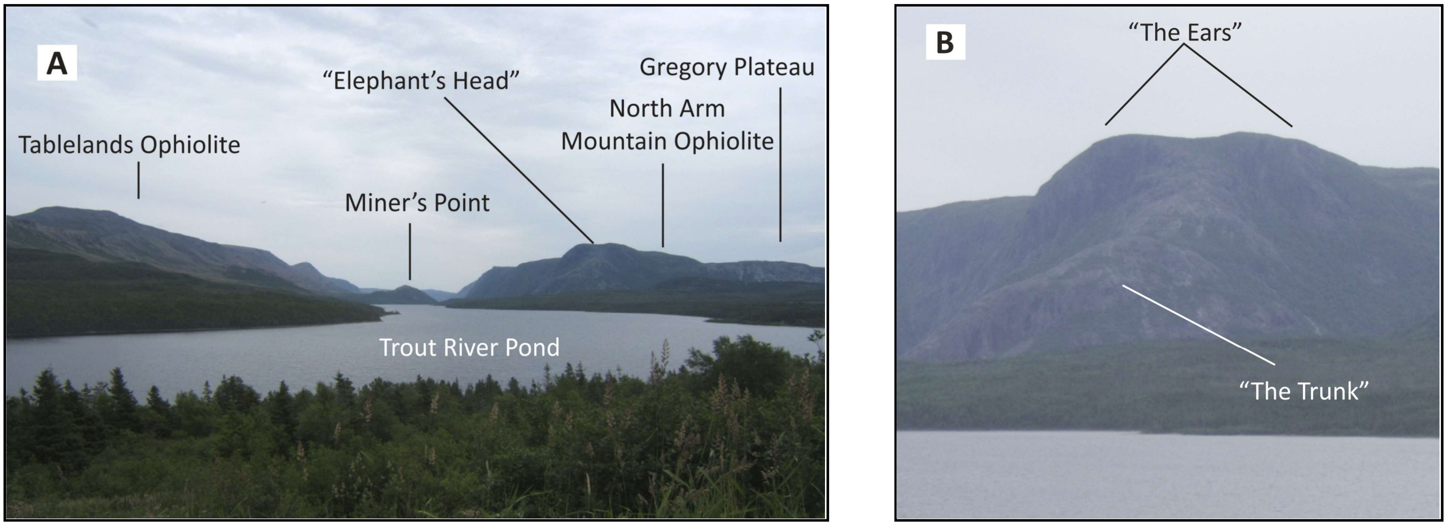

Figure 24. The views from the location near Trout River Campground (Stop 3.1). (A) Trout River Pond separating the ultramafic rocks of the Tablelands (left) from the higher levels of the ophiolite sequence exposed in the North Arm Mountain massif (right). (B) A closer view of the mountain known as the "Elephant's Head" (officially termed Narrows Head) showing its resemblance to the folded ears and long trunk of an elephant. Most of this prominent feature consists of sheeted diabase dykes.

from a distance, but the boat tour availability has varied in recent years. Excursion 3 is a hike of about 11 km roundtrip, which starts at the boat launching area, and follows an official Park trail; it is a relatively easy walk, without any significant elevation gain, and is well marked. The hiking trail encounters little outcrop except at its easternmost end, but it provides access to peridotites and gabbros of the Tablelands Ophiolite without actually climbing to the high plateau. The trail also provides many scenic views of the Tablelands and North Arm Mountain, including distant views of the sheeted dykes in the latter. At least 4–5 hours should be allowed for the return hike. A visit to the scenic viewpoint by the campground (Stop 3.1; signposted) is also highly recommended.

Stop 3.1: Viewpoint Near the Trout River Campground (418521E, 5479170N)

If entering Trout River from Route 431 westbound, turn left at the junction, and continue west beside the river, following the signs for the campground. The road towards the campground is the extension of Main Street in the village. Turn right at the junction for the day-use area, cross the river bridge, and ascend to the viewpoint, signposted on the right.

This viewpoint gives one of the more spectacular views in the Park (Fig. 24), and it contains a wealth of geological information. The south side of the Tablelands Ophiolite is clearly visible, and the contrast between red-weathering ultramafic rocks and grey-weathering gabbro is obvious. Miner's Point, at the head of the lake, also consists of gabbro. A large detached piece of gabbro has moved downslope towards the lake in postglacial times; this is visited at a later stop on the hike. A vertical fault occupies the lake valley, and the hills to the south (right) are part of the North Arm Mountain massif. This area is dropped down relative to the Tablelands Ophiolite and also tilted towards the viewpoint, exposing the metamorphic aureole of the ophiolite at the far end of the lake. The hills on the south side of the lake thus expose higher sections of the oceanic crust, such as sheeted dykes and gabbro, which are less abundant in the Tablelands. A prominent rounded peak has a long ridge that connects its summit to the shore of the lake, and rounded cliffs on either side (Fig. 24). This is known locally as the "Elephant's Head", because the long ridge resembles a trunk and the rounded cliffs resemble ears. It consists mostly of sheeted diabase dykes. The high peak in the south is Mount St. Gregory (674 m), which is one of the highest points in Newfoundland. The Mount St. Gregory area was well-known for numerous but unfortunately small copper-bearing veins, which is how Miner's Point got its name.

Stop 3.2: Mélange and Shale Outcrops (420623E, 5477479N)

From the viewpoint (Stop 3.1) return towards Route 431, turn right after crossing the bridge, and then take the side road signposted for the day-use area. The trail starts just behind the toilets. There are no outcrops at all for the first 2.5 km, but the trail is very pleasant, running through thick woodland just above the shore of the lake. The lack of outcrop in the first part of the trail reflects the fact that it is largely within the mélange unit that sits beneath the Tablelands Ophiolite. Some reddish shales of this unit are exposed within a small stream at this location, but may not always be visible above the water level.

Stop 3.3: Stream Crossing and Views of Slumped Gabbro Mass (420923E, 5477329N)

A few hundred metres beyond Stop 3.2, the trail emerges into countryside that is more open and crosses a larger brook. The boulders in the brook include numerous peridotites and lineated to foliated gabbros. The latter resemble the gabbros that outcrop in the region of the Moho (see Excursion 4) and may be derived from similar sources.

This location has a view of huge grey outcrop that has a flat top, sloping to the south (Fig. 25). The grey outcrop con-

Figure 25. The large mass of gabbro (grey) sitting physically above peridotite (orange-brown) in the valley of Trout River Pond (Stop 3.3). This feature was interpreted by Smith (1958) to be a product of postglacial mass movements in the valley and its glacially oversteepened cliffs.

sists of gabbro, whereas the outcrops beneath it are clearly peridotite, and they have the distinctive red-brown colour of the Tablelands. However, the contact between these two rock types is not the Moho in this particular location. Smith (1958) proposed that this enormous crag of gabbro "slumped" into the valley of Trout River Pond in postglacial times, when the support of the ice sheets was lost following their melting. Large areas of slumping are identified by airphoto analysis of the areas, and smaller areas of gabbro scattered through the valley sides were interpreted in a similar way (Smith 1958). Further discussion of this idea is provided by Brookes and Deardon (1981) and Brookes (1993).

Stop 3.4: Gabbro Outcrops (422073E, 5476584N)

Stop 3.4 is located about 4.5 km from the trailhead. There are 'outcrops' immediately above the trail that belong to the slumped gabbro mass, and large blocks adjacent to the trail that represent the same material. The rock type is an altered and retrogressed foliated gabbro, which contains some discontinuous lenses of ultramafic composition, possibly feldspathic dunites. The gabbros are not as well-preserved as those found on the Tablelands high country excursion (see Excursion 4), but they are likely derived from the region of the Moho.

This location also affords excellent views of the narrows in Trout River Pond, and the imposing cliffs on the opposite side of the lake (Fig. 26). The highest cliff visible across the lake (part of the "Elephant's Head" noted at Stop 3.1) is mostly formed by sheeted diabase dykes, representing the feeder systems to seafloor mafic volcanic rocks. The lowermost part of the cliff consists of gabbro. The vertical dykes in the cliff face are just visible to the naked eye if you have keen eyes, but are more easily seen through binoculars.

From Stop 3.4 onwards, the trail passes through ultramafic rocks as evidenced by the barren country and numerous peridotite boulders. Note the view down to the beaches on either side of the narrows in Trout River Pond; these display the contrasts in geology very nicely. The beach on the north side is orange, as it is largely of ultramafic composition, whereas the beach on the south side is grey, being dominantly mafic (Fig. 26).

Stop 3.5: Peridotites (422943E, 5476064N)

The outcrops at Stop 3.5 lie very close to the end of the trail, and are the most extensive areas of bedrock anywhere on the trail. They consist of brown-weathering peridotite and harzburgite, cut by both pyroxenite and dunite dykes (Fig. 27). The rocks contain a weakly developed fabric. Looking up towards the cliffs of the Tablelands Ophiolite, you may see several greyish patches indicating the locations of alkaline springs, similar to the one noted in Winterhouse Brook on Excursion 2. There are also very good views of the Trout River Pond valley and Gregory Plateau from this location.

EXCURSION 4: THE LOWER OCEANIC CRUST, THE MOHO AND THE MANTLE ON THE TABLELANDS SUMMIT PLATEAU

General Information

Excursion 4 is a lengthy off-trail hike, but well worth the effort. It visits the Moho, which marks the gradational transition between ultramafic rocks of the mantle and gabbroic rocks of the lower section of the oceanic crust. Early studies of the Bay of Islands Igneous Complex (BOIC) called this the "Critical Zone" (Smith 1958), and this term is retained in some accounts. This terminology is derived from large layered intrusions such as the Bushveld Complex of southern Africa, where it refers to the region where cumulus plagioclase first appears in the crystallization sequence. The hike initially crosses through gabbroic rocks of the lowermost oceanic crust and, if completed fully, eventually traverses some of the spectacular scenery formed by ultramafic rocks on the high plateau.

Specific safety concerns apply to this excursion. It should not be attempted unless you are in good physical condition and well prepared. It should never be attempted if the summit plateau is covered in cloud, and hikers should quickly descend if the weather deteriorates. The time required depends on how many of the stops below are included. A straight hike to the Moho and back on the same route can be done in about 5 hours, but if the return loop through the underlying ultramafic rocks is included, at least 7 or 8 hours should be allocated. Although it is quicker to retrace your steps from the Moho, the walk across the ultramafic plateau is a rather surreal experience. The area around the Moho is also described by Malpas (1987), and information is also provided by Stevens et al. (2003), but without locational and directional information.

Climbing from Route 431 to Stop 4.1

The starting point for the hike is the parking lot for the Green Gardens Trail on Route 431, between the Discovery Centre in Woody Point and Trout River. The UTM coordinate for this spot is approximately 424650E, 5283050N. Note that there are two separate trailheads for the Green Gardens Trail system, and that this is the first parking area on the right hand side when heading west on Route 431 towards Trout River. If trav-

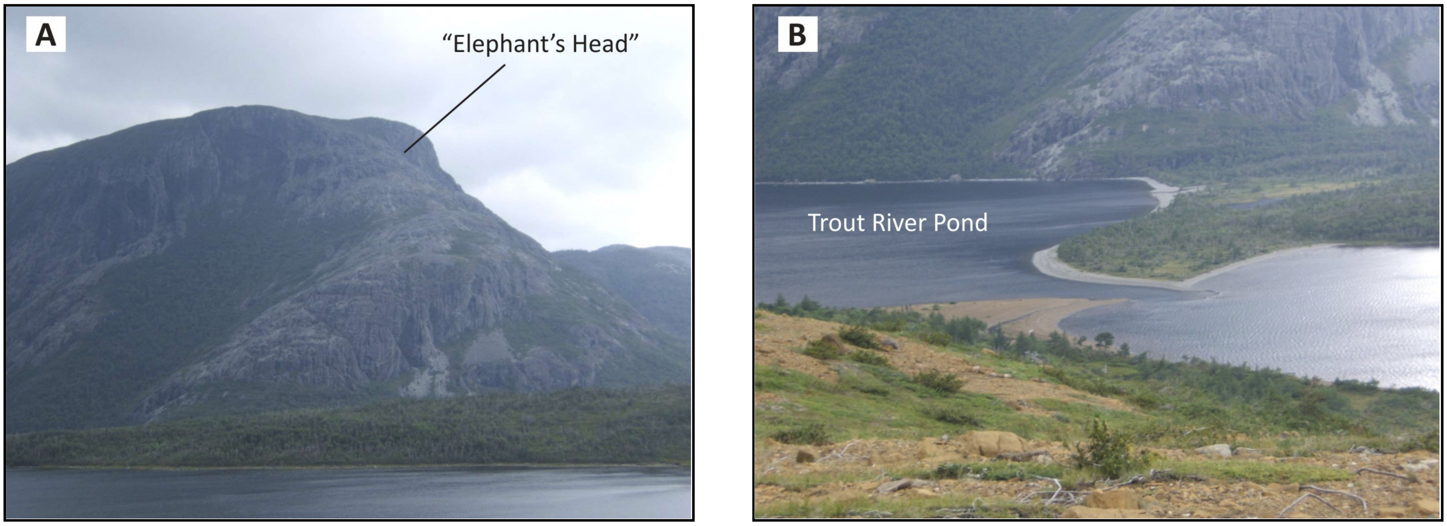

Figure 26. Views from the Trout River Pond Hiking Trail, at Stop 3.5. (A) Another view of the "Elephant's Head", where sheeted diabase dykes can be seen by those with keen eyes or good binoculars. (B) The contrasting colours of cobble beaches on either side of Trout River Pond Narrows that reveal the contrast between peridotite (foreground) and gabbro or diabase (background).

elling east, it is the second parking lot on the left. In 2019, it was not visibly identified for the Green Gardens Trail, due to flood-related damage on parts of the trail. From the trailhead parking lot, walk a short distance west, and then leave the road to the left just past the large signboard. Ascend the steep slope on the west bank of a small stream.

Stop 4.1: Route Orientation (424448E, 5482789N)

There is no outcrop at this location, but it is a good place to take a rest. The site consists of a large pile of ultramafic boulders, which are probably frost-heaved blocks from underlying bedrock. From here, you can see your destination on the high ridge ahead. A prominent outcrop is visible just to the right of a notch where a small stream emerges. During the remainder of the steep ascent, keep this outcrop in sight and head directly towards it. You will probably have many chances to view the route ahead as you stop to regain your breath.

Stop 4.2: Coarse-Grained Gabbro of the Lower Oceanic Crust (424133E, 5482384N)

As you ascend the hill from stop 4.1 keep to the west (right hand side) of the stream valley. Note that the brown ultramafic talus becomes mixed with grey blocks of gabbroic composition as the hill is climbed. There are few definite outcrops enroute, because the surface is extensively frost-shattered, but the transition between the ultramafic rocks and the gabbro is marked, as elsewhere in the park, by increasing amounts of scrubby vegetation.

Stop 4.2 is the first large outcrop on the route, and it consists of coarse-grained gabbro. The contact between the ultramafic rocks of the lower slope and this gabbro is a later fault zone, and does not actually represent the Moho at this location. It is hard to see fresh surfaces in this outcrop. The gabbro is medium- to coarse-grained and consists largely of plagioclase and subophitic clinopyroxene, with lesser orthopyroxene. There is some local alteration of the mafic minerals to amphibole and chlorite. The absence of primary hydrous minerals and the low modal abundance of orthopyroxene suggest that these rocks represent an intermediate level in layer 3 of the oceanic crust. There are also some local concentrations of diabase dykes in this general area (but not seen at this outcrop) that probably represent the roots of the overlying layer 2 (sheeted dykes) which has largely been eroded.

Figure 27. The visually striking peridotite-dominated landscape around the end of the Trout River Pond Hiking Trail (Stop 3.5). The upper, more remote, part of Trout River Pond is visible in the distance.

Stop 4.3: Coarse-Grained Gabbro (424318E, 5481889N)

From Stop 4.2, make a slight change of direction to southeast as you continue up the slope. This will eventually lead to the large outcrops of Stop 4.3, which is the prominent rocky hill first observed from Stop 4.1, and almost on the summit plateau. At this point, take a well-deserved rest and admire the wonderful views of Gros Morne mountain to the northeast, and the broad sweep of the coast directly to the north, beyond Rocky Harbour.

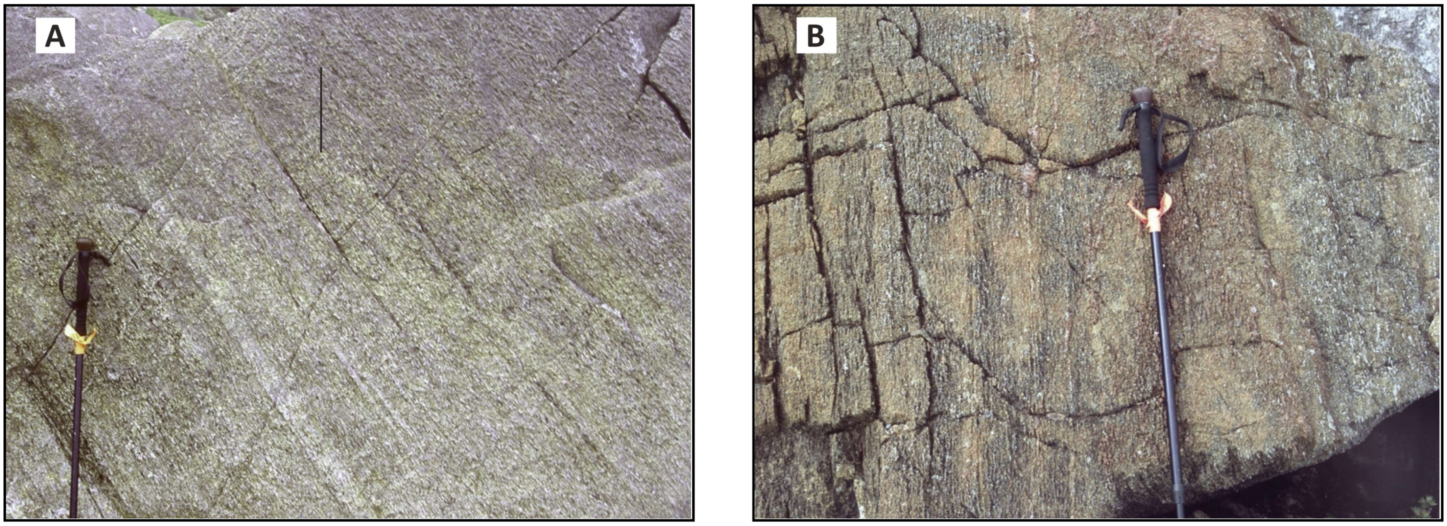

Figure 28. Examples of gabbroic rocks seen on the ascent to the Tablelands plateau (Stops 4.3 and 4.5). (A) Foliated gabbro, with prominent leucocratic bands. (B) Mafic cumulate rocks with brownish troctolitic intervals and grey gabbroic material; it is not clear if the foliation is magmatic or superimposed.

The outcrop at Stop 4.3 consists of coarse-grained gabbro that resembles the outcrops at Stop 4.2, but it is much better preserved. It contains a well-developed plagioclase alignment and foliation that at first sight appears to be of primary magmatic origin (Fig. 28). Some compositional layering is developed, but this is generally better seen in loose blocks than in the outcrops. Petrographic studies indicate that the layering and foliation are partly due to deformation and recrystallization under high-temperature ductile conditions rather than magmatic accumulation (Calon et al. 1988; Stevens et al. 2003). However, it is likely that this recrystallization and flow closely followed the crystallization of the host rocks, as these fabrics are locally cut by undeformed gabbro and anorthositic segregations.

Stop 4.4: Views of the Plateau and Moho Region (424463E, 5481814N)

From Stop 4.3, continue for about 150 metres to the east, to the true summit of the hill. The outcrops here are essentially the same as those described from Stop 4.3, but more extensive. From here, the next section of the hike is clearly visible. To the east, the Tablelands summit plateau is now visible, consisting largely of brown-weathering ultramafic rocks; these are incised by a shallow stream valley. To the right of the stream valley is a large barren hill with brown ultramafic rocks showing through a ragged cover of green grass, which is in turn flanked to the right by a more vegetated hill on which the rocks are grey in color. This hill is located on a bearing of approximately 120°, and consists of gabbro; the Moho separates these mafic rocks from the ultramafic rocks to the east. This grey hill represents Stop 4.5, and the route heads directly towards it.

Stop 4.5: Strongly Foliated and Lineated Gabbro (424958E, 5480504N)

From Stop 4.4, walk south-southeast for approximately 1.3 km to reach the north end of the long ridge that exposes the Moho. The terrain is generally flat, and the walking is for the most part relatively easy, albeit swampy in places. Some sections consist of frost-heaved and broken rock that is hard on the ankles and requires some caution. However, this is nothing compared to the broken terrain that you will encounter later when you get into the ultramafic rocks in the final section of the hike.

Stop 4.5 is a prominent grey outcrop with many large frost-heaved boulders. The hike from Route 431 to this spot will generally take about 2 hours, and this is one of two places recommended for a possible lunch break. The second location is Stop 4.6, only 200 m south from which the views are superior, but likely to be windier. Stop 4.5 provides better shelter from the wind in the lee of some of the large blocks.

Outcrops at Stop 4.5 consist of coarse-grained, foliated to lineated gabbro, showing variably-developed compositional layering (Fig. 29). However, there are also many blocks of feldspathic dunite, which are typically chocolate-brown weathering, and contrast with the grey-weathering gabbro. South of Stop 4.5 towards the crest of the ridge, there are several outcrops of this material, many of which contain fine-grained plagioclase-rich (nearly anorthositic) layers. Minor chromite is visible on many weathered surfaces, and chromite-rich pods are locally visible, but there are no significant chromite accumulations. This is a gradational contact region, with feldspathic dunites to the east, gabbro to the west, and interlayered dunites and gabbros in the middle. It represents the geophysical Moho, despite the fact that there is an obvious contrast in petrology; the compositional shift from gabbro to peridotite would result in an increase in seismic velocity. In essence, the Moho at this site is a strongly deformed package of ultramafic and mafic rocks including dunite, felspathic dunite (described as troctolite by Malpas 1987), pyroxene-rich gabbro and olivine gabbro, and a distinctive assemblage of anorthositic dykes and veins. The strong fabrics and numerous folds indicate intense plastic (ductile) deformation, and several features (notably rotated clinopyroxene megacrysts) indicate that the Moho was essentially a large-scale shear zone. The fabrics and the evidence of

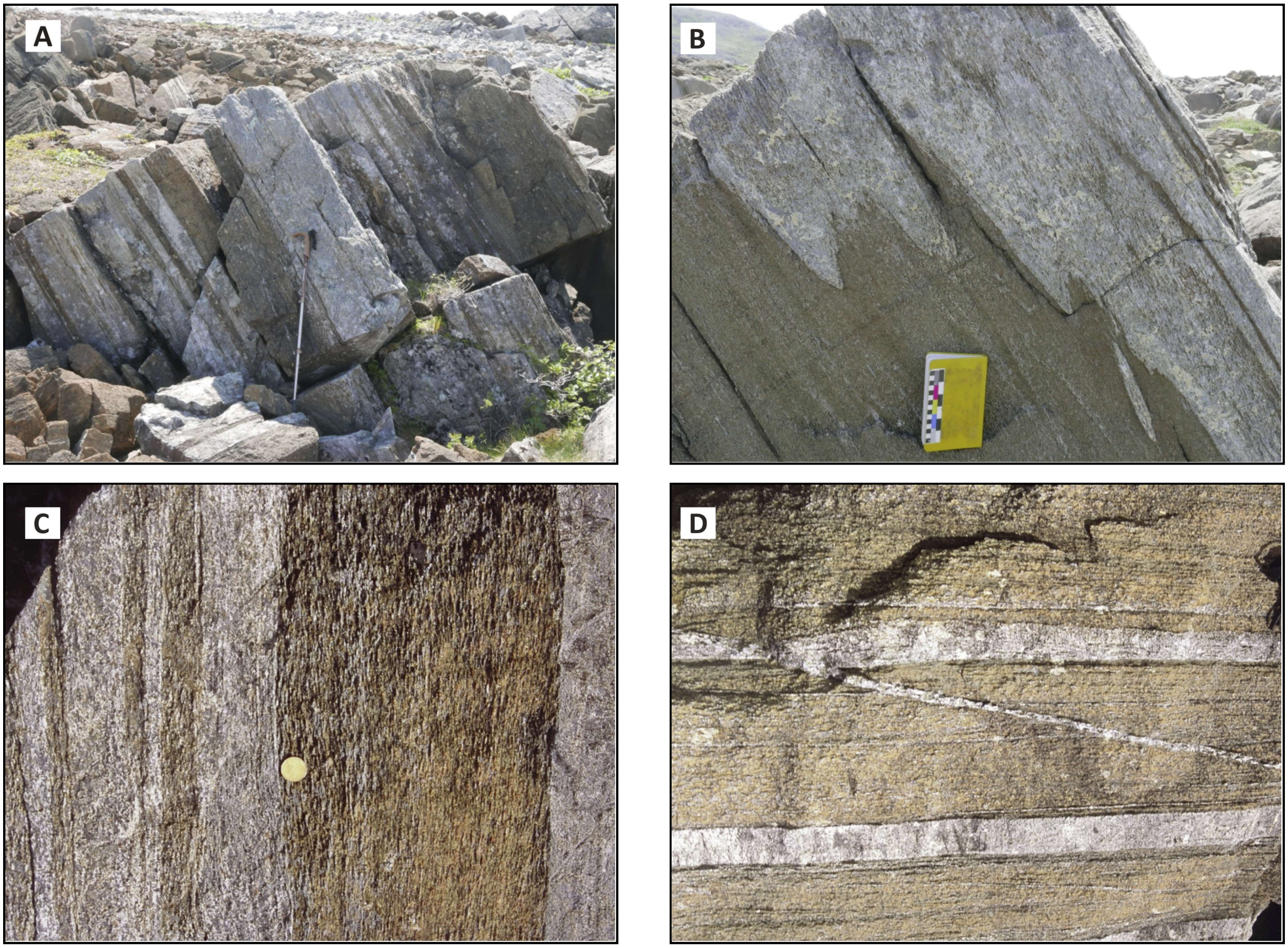

Figure 29. Examples of rock types seen in the region of the Moho at Stop 4.6 and adjacent areas. (A) Typical outcrops showing the mixed character of the Moho region, with alternating zones of variably layered gabbro (grey) and olivine-rich ultramafic rocks (brown); in detail, such compositional domains are more like elongate lenses than layers. (B) Leucocratic layers, preserving contact relationships with associated mafic cumulate rocks. (C) Strongly foliated feldspathic dunite showing colour variations defined by the plagioclase to olivine ratio. (D) Sharply-defined plagioclase-rich bands (originally cumulates?) in a rock type similar to (C), but note also the cross-cutting plagioclase-rich vein in the centre of the photo. Photos (A) and (B) by John Waldron, University of Alberta.

deformation are in many respects similar to the features observed near Winterhouse Brook (Excursion 2) and in later parts of this excursion. Deformation of this intensity and type is almost never seen in the layered sequences of large mafic intrusions such as the Bushveld Complex, and its presence puzzled earlier geologists who interpreted the BOIC from this perspective.

Stop 4.6: Moho Region (424893E, 5480329N)

Stop 4.6 is located on the ridge crest about 200 m south of Stop 4.5, and is perhaps the best spot to view the Moho and to photograph the location with spectacular views in the background (Fig. 30). The contrast between grey gabbros to the west and brown dunites to the east is very clear here. There are numerous white anorthositic layers in the dunite; some of these are discordant to the foliation (Fig. 29), implying that they represent originally cross-cutting veins that were later rotated into parallelism.

Stop 4.7: Deformed Augite Megacrysts (424858E, 5480379N)

Stop 4.7 is the grey layered outcrop that is visible from Stop 4.6, and located about 70 m to the northwest. It is a gabbroic outcrop that is notable for the presence of strong layering, foliation and many olivine-rich layers. It also contains some large clinopyroxene (augite) megacrysts, which have become augen due to pervasive shearing (Fig. 31). The outcrop also contains some intrafolial isoclinal folds, which are much harder to see than the megacrysts.

Figure 30. Panoramic view of the Moho location on the Tablelands (near Stop 4.6), with pale grey gabbro in the background and interlayered mafic and ultramafic rocks in the foreground. The view in the far distance extends to the Lookout Hills and the north side of Bonne Bay.

Stop 4.8: Deformed Gabbro and Views to the South (424548E, 5479679N)

From Stop 4.7, walk east to the ridge crest again, and then walk south along the ridge. There is not very much outcrop enroute to Stop 4.8, and most of the ridge consists of frost-heaved rubble. One section is bizarre in that there appears to be east-trending zones of gabbroic and ultramafic rubble, which are at right angles to the regional trend of the contact between these two rock types. It seems unlikely that this actually reflects the distribution of units in the bedrock, and perhaps the pale grey gabbro has been locally transported eastward by glacial processes. Alternatively, perhaps there are larger-scale fold structures in this region, which are difficult to see because the terrain is so broken.

Stop 4.8 consists of more continuous outcrop of deformed gabbro, also containing augen-like clinopyroxene (augite) megacrysts akin to those described at Stop 4.7. However, deformation appears to be much stronger here than at Stops 4.6 and 4.7, and there are several areas of streaky, finely banded gabbro that have a distinctly mylonitic appearance. These fabrics likely result from intense transposition of interlayered gabbro and dunite.

Figure 31. Large rotated clinopyroxene megacryst in strongly layered and foliated gabbro within the Moho region (Stop 4.7). This same feature was illustrated in the article by Malpas (1987).

Figure 32. Folds defined by leucocratic layers within the mafic–ultramafic cumulate sequence of the Moho region at Stop 4.9.

Stop 4.8 also reveals the first views to the south and southwest. The high hills to the south are part of the Gregory Plateau, consisting of the higher levels of the ophiolite suite, predominantly sheeted dykes and pillow lavas. These rock units are located on the other side of Trout River Pond, across a later fault, and are part of the North Arm Massif of the BOIC. (Further information is provided in the descriptions for Excursion 3). The valley to the southwest is occupied by the lower part of Trout River Pond, and lies mostly within mélanges that sit beneath the ultramafic rocks.

Stop 4.9: Fold Structures in the Moho Region (424488E, 5479579N)

Stop 4.9 is located a short distance south of Stop 4.8, and displays similar rock types and panoramic views. It also displays some folds developed within layered gabbro (Fig. 32). Some of the folds appear to be intrafolial structures, i.e. early folds that have been completely disrupted by later deformation. This is part of the process of transposition, which creates the strongly banded rocks seen at Stop 4.8 and elsewhere.

Stop 4.10: Transition Zone Beneath the Moho (424617E, 5479079N)

Stop 4.10 is a prominent grey area visible downhill from Stop 4.9. The walk from Stop 4.9 to Stop 4.10 passes back through the Moho again, and shows the same transition from layered foliated gabbros containing olivine-rich layers into feldspathic dunites containing plagioclase-rich layers. Towards Stop 4.10, there are also many blocks of harzburgite derived from the east, and the contrast between their yellow-brown weathering and the dark-brown weathering of the dunites becomes more obvious. There is no outcrop at Stop 4.10, which is dominated by large frost-heaved blocks of harzburgite.

The dunitic and harzburgitic rocks that sit beneath the Moho are part of the transition zone, which is believed to be derived from the cumulate crystal mush that developed at the base of the mafic magma chamber at the original spreading centre. The lower parts of the transition zone consist largely of harzburgites that contain pods and dykes of dunitic and pyroxenitic composition. These are the same rock types that are exposed in the area around Winterhouse Brook (Excursion 2).

This is the turnaround point for the shorter, easier version of Excursion 4. If you do not wish to descend into the rocks of the upper mantle, simply retrace the route described above. The next section of the full hike involves some difficult walking in broken terrain; the smooth brown mountains of ultramafic rock ahead of you look like easy walking, but distance is very deceptive, as much of the surface is chaotically broken.

Stop 4.11: Harzburgite Containing Pyroxenitic and Dunitic Veins (425143E, 5479144N)

From Stop 4.10, the route heads almost due east, downhill at first, across a small stream, and then up a gentle slope on the other side. This walk goes from the dunitic rocks that lie east of the geophysical Moho and into the yellow-brown weathering harzburgites that dominate most of the Tablelands plateau. The petrological Moho, as distinct from the geophysical Moho, is crossed somewhere in this interval, but is a much more subtle feature because it is manifested largely by textures in the ultramafic rocks, rather than being an obvious compositional boundary.

Outcrops at Stop 4.11 consist of coarse-grained harzburgite cut by two generations of younger veins. Fine-grained, brown-weathering veins appear to be of dunitic composition, and are themselves cut by coarse-grained, pale green, pyroxenitic veins. The relationships of these veins to the host harzburgites resemble those observed near Winterhouse Brook (Excursion 2) but here there appears to be little or no superimposed deformation.

Stop 4.12: Harzburgite Hill (425963E, 5479954N)

Stop 4.12, located on the summit of a broad hill, consists entirely of harzburgites. It is almost 1 km from Stop 4.11 to Stop 4.12, and the walk is not exactly pleasant, as most of the route consists of large frost-heaved harzburgite blocks. This type of material is typical of much of the plateau on top of the Tablelands Ophiolite, which looks quite inviting from a distance. However, the landscape is surreal and extraterrestrial in appearance. The outcrops between Stop 4.11 and Stop 4.12 are all harzburgites, essentially identical to those at Stop 4.11. The hilltop is worth exploring for its views. This is the highest point on the hike (just over 650 m), but not the highest point in the Tablelands Ophiolite, which is located at 721 m, east of the head of Winterhouse Brook canyon. The highest point is only about 2 km southeast of here, but is broad and flat-topped, with inferior views, and probably not worth the grueling walk across the broken landscape. Table Mountain itself is a smaller peak of gabbro seen to the west northwest, but its elevation is a mere 520 m.

Stop 4.13: Canyon Access Point (425458E, 5481144N)

From Stop 4.12, head northwest across the seemingly endless plain of frost-heaved harzburgite to this spot, located above a small canyon that descends the northern slope of Table Mountain. This canyon is also rather extraterrestrial and lifeless in appearance, totally lacking the vegetation that normally marks a watercourse. The canyon walls are loose scree slopes, and must be treated with caution. However, it is possible to descend into the canyon and up the opposite side, with care. After doing so, follow the stream for a few hundred metres. The walking by the stream is difficult, but most of the blocks are stable. Where the gradient of the stream suddenly increases, leave the valley again and follow the contours of the hillside to the northwest. From here, you can see Route 431 below, and can walk towards it directly. The route down is relatively easy if you manage to stay away from several areas dominated by large frost-heaved blocks. Upon reaching the road, walk westwards for about 1 km to return to the parking area (UTM coordinate 424650E, 5283050N) where the hiking route commenced.

ACKNOWLEDGEMENTS

This article, although long, is but a small sample of the scenic and geological attractions of Gros Morne National Park. Much of it started as part of a larger project initiated around 2001 to write a geological guidebook to the Park for visitors, in conjunction with the late Robert K. Stevens. Like many overambitious ideas, it was only partly completed, spending more than a decade in a filing cabinet and on obsolete floppy disks. Adapting this section for Geoscience Canada was far more time-consuming than I ever anticipated, but hopefully other parts of this project may yet see the light of day. Robert (Bob) Stevens played a crucial role in understanding the geology of western Newfoundland, and was the first to correctly assemble all the pieces of the Gros Morne jigsaw puzzle. His work focused on the sedimentary rocks and their fossils, but he noted many other things on many other topics, often preferring new questions to firm answers. His contribution to knowledge is not fully reflected in the number of publications that he authored. Bob was not the one who first introduced me to Gros Morne National Park, but it was through knowing and talking with Bob that I gained a better but still incomplete understanding of this amazing place.

I also acknowledge many useful general and detailed comments from reviewers John Waldron and Peter Cawood, which greatly improved a rather sloppy first draft. Brendan Murphy is also thanked for editorial handling of the paper on behalf of Geoscience Canada.

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Received May 2019
Accepted as revised August 2019

APPENDIX: GLOSSARY OF SELECTED TERMS

The following is included to assist readers with limited knowledge of geological terms beyond basic concepts. It is not exhaustive, but any undefined terms in the text can probably be understood with reference to one of several popular geoscience directories.

Accretion (tectonic): a large-scale process in which continental fragments and island arcs are swept in towards larger continents by plate tectonics, and then joined to them. Characteristic of *convergent margins*.

Allochthon, allochthonous: packages of rock units that formed in another place but were then transported by the forces of plate tectonics to the location in which they are now found. Opposite of *autochthonous*.

Anorthositic, anorthosite: igneous rock consisting largely of Ca–Na-rich feldspar (plagioclase) with very small amounts of olivine and/or pyroxene. Often a *cumulate rock*.

Autochthon, autochthonous: packages of rock units found in essentially the same location where they originally formed. Opposite of *allochthonous*.

Convergent margin: a location where tectonic plates are moving toward each other and one plate descends beneath the other by the process of *subduction*.

Cumulate rock: igneous rock formed by the accumulation of minerals that have crystallized from a liquid magma. See also *fractionation*.

Diachronous: adjective describing a process or event that occurs at different times in different places, but often in a rather orderly progressive fashion. For example, the human migration out of Africa to other continents was diachronous. It is the opposite of synchronous.

Dunite: a type of peridotite consisting mostly of the mineral olivine.

Fractionation: the process by which a liquid magma changes composition as different minerals crystallize and separate from the remaining liquid magma by some process of accumulation, to make a cumulate rock.

Global Stratotype Section: the location where a geologic boundary is defined between rock layers or with reference to fossil assemblages in the rock layers, as a reference point for the geological time scale.

Gondwanaland: a large Paleozoic continental assembly including much of Africa, South America, Australia and Antarctica, along with parts of Asia. The Appalachian–Caledonian Orogenic Belt was created as part of the process that joined Gondwanaland to North America to create the supercontinent called Pangaea ('all land') at the end of the Paleozoic Era.

Harzburgite: a type of peridotite consisting mostly of the mineral olivine and orthopyroxene, which is a Mg–Fe-rich pyroxene.

Hydrothermal: processes involving hot waters and aqueous solutions circulating in Earth's subsurface.

Intrafolial fold: an isoclinal fold that is disconnected from adjoining layers; normally an indication of a larger and older fold structure that has been torn apart by continued deformation.

Isoclinal fold: a tight fold structure in which the two fold limbs are essentially parallel; a common product of intense deformation under ductile conditions.

Mélange: a mixed up rock, usually deformed and badly fractured, in which blocks of different rock types of varied sizes are contained within a matrix of sedimentary origin, usually in a rather chaotic fashion.

Metamorphic aureoles: zones of metamorphic rocks produced through thermal effects related to the proximity of high-temperature liquid magmas, or hot solid rocks such as mantle peridotites.

Metasomatism: a process by which the bulk composition of a rock unit is changed, normally through reactions with hydrothermal fluids percolating through it.

Mohorovičić Discontinuity (Moho): a region in the crust where geophysical studies indicate an increase in the velocity of seismic waves, consistent with a sudden increase in the density of rock units at or near the boundary between the Earth's crust and the upper mantle region.

Mylonite zone: a belt of fine-grained and intensely banded rocks that geologists usually consider to indicate relative motion of adjoining rock units under high-temperature conditions that lead to ductile behaviour of minerals in the area caught between those rock units.

Obduction: process that places rocks of the oceanic crust above those of the continental crust; the opposite of *subduction*.

Paleoproterozoic: a part of the vast Precambrian time period, extending from 2500 to 1600 million years ago.

Partial melting: process by which some of the minerals in a silicate rock melt due to high temperatures, creating a liquid magma that has a different composition to the starting material and the residual material left behind. See also *fractionation*.

Peridotite: general term for ultramafic igneous rocks consisting mostly of the minerals olivine and various pyroxenes.

Pillow lava: a submarine volcanic rock in which basalt magma chills rapidly into blob-like structures, much as hot honey would congeal on contact with ice-cold water.

Pyroxenite: a rock consisting largely of pyroxene minerals; orthopyroxene and clinopyroxene minerals may occur alone or in combination.

Retrogressed, retrogression: changes in the mineral assemblage found in an igneous rock to water-bearing minerals characteristic of lower temperatures of formation, but without any significant change in the rock's bulk composition.

Serpentinites, serpentinization: rocks that consist largely of serpentine-group minerals, which are soft flaky silicate minerals that resemble snakeskins when weathered. Serpentine minerals are produced by reactions between primary igneous minerals in ultramafic rocks and percolating waters or *hydrothermal* fluids.

Subduction: when one tectonic plate (usually oceanic) descends beneath another, usually resulting in earthquakes and abundant volcanic activity. A characteristic process at *convergent margins*.

Syncline, synclinal: a gentle bowl-like fold structure of rock units, such as thinly-bedded sedimentary rocks or km-scale units of components in an ophiolite complex.

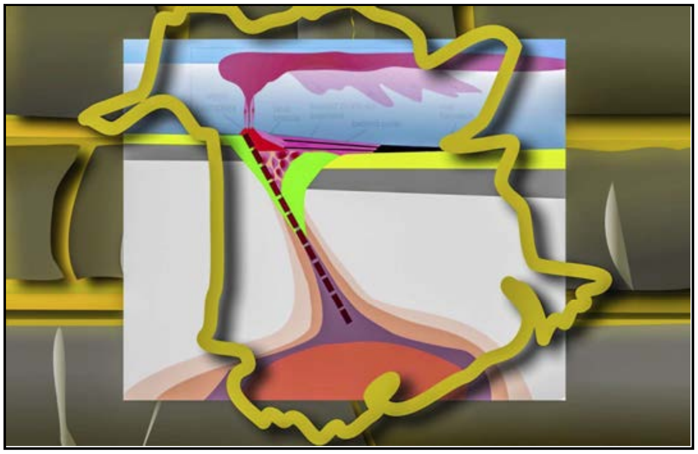
Tethys Ocean: a largely vanished ocean, which once separated Africa and parts of Asia from Europe. The Mediterranean sea is all that now remains of the Tethys Ocean.

Transposition: a process that creates intensely banded rocks such as mylonites, or the banded ultramafic rocks described from the mantle regions of ophiolite complexes.

Trondhjemite: an unusual type of igneous rock that resembles a granite, but consists largely of quartz and the Na-rich feldspar mineral albite (most granites also contain K-rich feldspar minerals such as orthoclase or microcline).

Wehrlite: a type of peridotite consisting mostly of the mineral olivine and clinopyroxene, which is a Ca–Mg-rich pyroxene variety.

SERIES



Great Mining Camps of Canada 7. The Bathurst Mining Camp, New Brunswick, Part 1: Geology and Exploration History

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SUMMARY

The Bathurst Mining Camp of northern New Brunswick is approximately 3800 km^2 in area, encompassed by a circle of radius 35 km. It is known worldwide for its volcanogenic massive sulphide deposits, especially for the Brunswick No. 12 Mine, which was in production from 1964 to 2013. The camp was born in October of 1952, with the discovery of the Brunswick No. 6 deposit, and this sparked a staking rush with more hectares claimed in the province than at any time since.

In 1952, little was known about the geology of the Bathurst Mining Camp or the depositional settings of its mineral deposits, because access was poor and the area was largely forest covered. We have learned a lot since that time. The camp was glaciated during the last ice age and various ice-flow directions are reflected on the physiographic map of the area. Despite abundant glacial deposits, we now know that the camp comprises several groups of Ordovician predominantly volcanic rocks, belonging to the Dunnage Zone, which overlie older sedimentary rocks belonging to the Gander Zone. The volcanic rocks formed during rifting of a submarine volcanic arc on the continental margin of Ganderia, ultimately leading to the formation of a Sea of Japan-style basin that is referred to as the Tetagouche-Exploits back-arc basin. The massive sulphide deposits are mostly associated with early-stage, felsic volcanic rocks and formed during the Middle Ordovician upon or near the sea floor by precipitation from metalliferous fluids escaping from submarine hot springs.

The history of mineral exploration in the Bathurst Mining Camp can be divided into six periods: a) pre-1952, b) 1952-1958, c) 1959-1973, d) 1974-1988, and e) 1989-2000, over which time 45 massive sulphide deposits were discovered. Prior to 1952, only one deposit was known, but the efforts of three men, Patrick (Paddy) W. Meahan, Dr. William J. Wright, and Dr. Graham S. MacKenzie, focused attention on the mineral potential of northern New Brunswick, which led to the discovery of the Brunswick No. 6 deposit in October 1952. In the 1950s, 29 deposits were discovered, largely resulting from the application of airborne surveys, followed by ground geophysical methods. From 1959 to 1973, six deposits were discovered, mostly satellite bodies to known deposits. From 1974 to 1988, five deposits were found, largely because of the application of new low-cost analytical and geophysical techniques. From 1989 to 2000, four more deposits were discovered; three were deep drilling targets but one was at surface.

RÉSUMÉ

Le camp minier de Bathurst, dans le nord du Nouveau-Brunswick, s'étend sur environ 3 800 km^2 à l'intérieur d'un cercle de 35 km de rayon. Il est connu dans le monde entier pour ses gisements de sulfures massifs volcanogènes, en particulier pour la mine Brunswick n° 12, exploitée de 1964 à 2013. Le camp est né en octobre 1952 avec la découverte du gisement Brunswick n° 6 et a suscité une ruée au jalonnement sans précédent avec le plus d'hectares revendiqués dans la province qu'à présent.

En 1952, on savait peu de choses sur la géologie du camp minier de Bathurst ou sur les conditions de déposition de ses gisements minéraux, car l'accès était très limité et la zone était en grande partie recouverte de forêt. Nous avons beaucoup appris depuis cette période. Le camp était recouvert de glace au

Geoscience Canada, v. 46, http://www.dx.doi.org/10.12789/geocanj.2019.46.150 pages 137–154

cours de la dernière période glaciaire et diverses directions d'écoulements glaciaires sont révélées sur la carte physiographique de la région. Malgré des dépôts glaciaires abondants, nous savons maintenant que le camp comprend plusieurs groupes de roches ordoviciennes à prédominance volcanique, appartenant à la zone Dunnage, qui recouvrent de plus vieilles roches sédimentaires de la zone Gander. Les roches volcaniques se sont formées lors du rifting d'un arc volcanique sous-marin sur la marge continentale de Ganderia, ce qui a finalement abouti à la formation d'un bassin de type mer du Japon, appelé bassin d'arrière-arc de Tetagouche-Exploits. Les gisements de sulfures massifs sont principalement associés aux roches volcaniques felsiques de stade précoce et se sont formés au cours de l'Ordovicien moyen sur ou proche du plancher océanique par la précipitation de fluides métallifères s'échappant de sources chaudes sous-marines.

L'histoire de l'exploration minière dans le camp minier de Bathurst peut être divisée en six périodes: a) antérieure à 1952, b) 1952-1958, c) 1959-1973, d) 1974-1988 et e) 1989-2000, au cours desquelles 45 dépôts de sulfures massifs ont été découverts. Avant 1952, un seul dépôt était connu, mais les efforts de trois hommes, Patrick (Paddy) W. Meahan, William J. Wright et Graham S. MacKenzie, ont attiré l'attention sur le potentiel minier du nord du Nouveau-Brunswick, ce qui a conduit à la découverte du gisement Brunswick n° 6 au mois d'octobre 1952. Dans les années 50, 29 gisements ont été découverts, résultant en grande partie de l'utilisation de levés aéroportés, suivis de campagnes géophysiques terrestres. De 1959 à 1973, six gisements ont été découverts. Ce sont essentiellement des formations satellites de gisements connus. De 1974 à 1988, cinq gisements ont été découverts, principalement grâce à l'utilisation de nouvelles techniques analytiques et géophysiques peu coûteuses. De 1989 à 2000, quatre autres gisements ont été découverts. Trois étaient des cibles de forage profondes, mais l'un était à la surface.

Traduit par la Traductrice

INTRODUCTION

The Bathurst Mining Camp (BMC), formerly referred to as the Bathurst – Newcastle Mining District, is known worldwide for its volcanogenic massive sulphide (VMS) deposits, especially for the Brunswick No. 12 Mine, which closed on April 30, 2013 after 49 years in operation. During its lifetime, this mine produced 136,643,367 tonnes of ore grading 3.44% Pb, 8.74% Zn, 0.37% Cu, 102.2 g/t Ag, making it one of the largest known and longest lived, underground VMS deposits in the world. This mine and the BMC contributed significantly to the economy of northern New Brunswick for nearly 50 years. It produced almost 500 million ounces of by-product silver (from lead concentrate) during its lifetime, making it one of the largest silver producers in North America. This paper describes the geological setting of the BMC and the history of exploration in this area up to closure of Brunswick No. 12.

Location and Physiography

The BMC of northern New Brunswick is mostly encompassed by a circle of radius 35 km (Fig. 1). The surface area is approximately 3800 km^2 but some of it is concealed by younger, shallow-dipping Carboniferous rocks to the east. The elevation ranges from approximately 50 m at Middle Landing in the east, to slightly over 600 m at Little Bald Mountain in the west, and elevation generally increases from east to west (Fig. 1). The topography consists of gently rolling hills, as a result of glacial erosion, with post-glacial, incised stream valleys. A good description of the physiography of the northern BMC can be found in Skinner (1974). Virtually all the BMC is forested. Much of it is covered by a thin (< 1 m) veneer of till but at lower elevations, till thickness increases and glacial-fluvial deposits are present. The area is drained by five major rivers and their tributaries: Middle and Tetagouche rivers in the northeast, Southeast Upsalquitch River in the northwest, Northwest Miramichi River (including the Sevogle River) in the south, and Nepisiguit River in the central part of the BMC. Provincial highways 180 and 430, and numerous logging roads currently provide easy access to the BMC, at least during the summer and autumn months, but most were not present when the BMC was discovered.

Camp Definition

In the September 10th, 1953 edition of the *Northern Miner*, is a map on page 5 of the "Bathurst Mining Area", which encompasses what we now consider to be the eastern part of the BMC. Then in the June 14th, 1956 edition of the *Northern Miner*, there is a lengthy article entitled: "*Hopeful tone to exploration in Bathurst-Newcastle Camp*", but no map showing the extent of this camp. The following year, in the April 18th edition of the *Northern Miner*, the technical program for the third day of the 59th CIM Annual Meeting in Ottawa is listed on page 24. The morning session of the Geology Division features a "*Symposium on Bathurst – Newcastle Area*", in which the first three speakers (C.H. Smith, G.S. MacKenzie, and S.H. Ward) refer to 'Bathurst – Newcastle Mining District' in their talk titles. In the July 1957 issue of the *Atlantic Advocate*, an article entitled: "*The Mineral Wealth of New Brunswick 1953–1957*" by G.S. MacKenzie contains a sketch map bearing the title "*Bathurst – Newcastle Mining District, New Brunswick*", which shows the locations of 28 known deposits. A technical paper by MacKenzie (1958) contained the same sketch map, but the limits of the district are spelled out in the text: "*The Bathurst – Newcastle district may be considered, for the purposes of this paper, to include the country from the town of Bathurst southward to Newcastle (now part of the city of Miramichi), westward to the headwaters of Nipisiquit* (sic) *river, and northwestward to the city of Campbellton*". Hence, the name, "Bathurst – Newcastle Mining District" is attributed to MacKenzie (1958).

Subsequently, the Bathurst – Newcastle Mining District was restricted to the belt of Cambro – Ordovician volcanic and sedimentary (now mostly Miramichi Group) rocks that were assigned to the original Tetagouche Group (Harley 1979), which extends southwestward from Bathurst. By the 1980s, the name had been shortened to Bathurst district or Bathurst camp (Franklin et al. 1981) but in the late 1990s, when a five-year, joint federal – provincial project (EXTECH-II) was conducted, the name Bathurst Mining Camp (BMC) became firm-

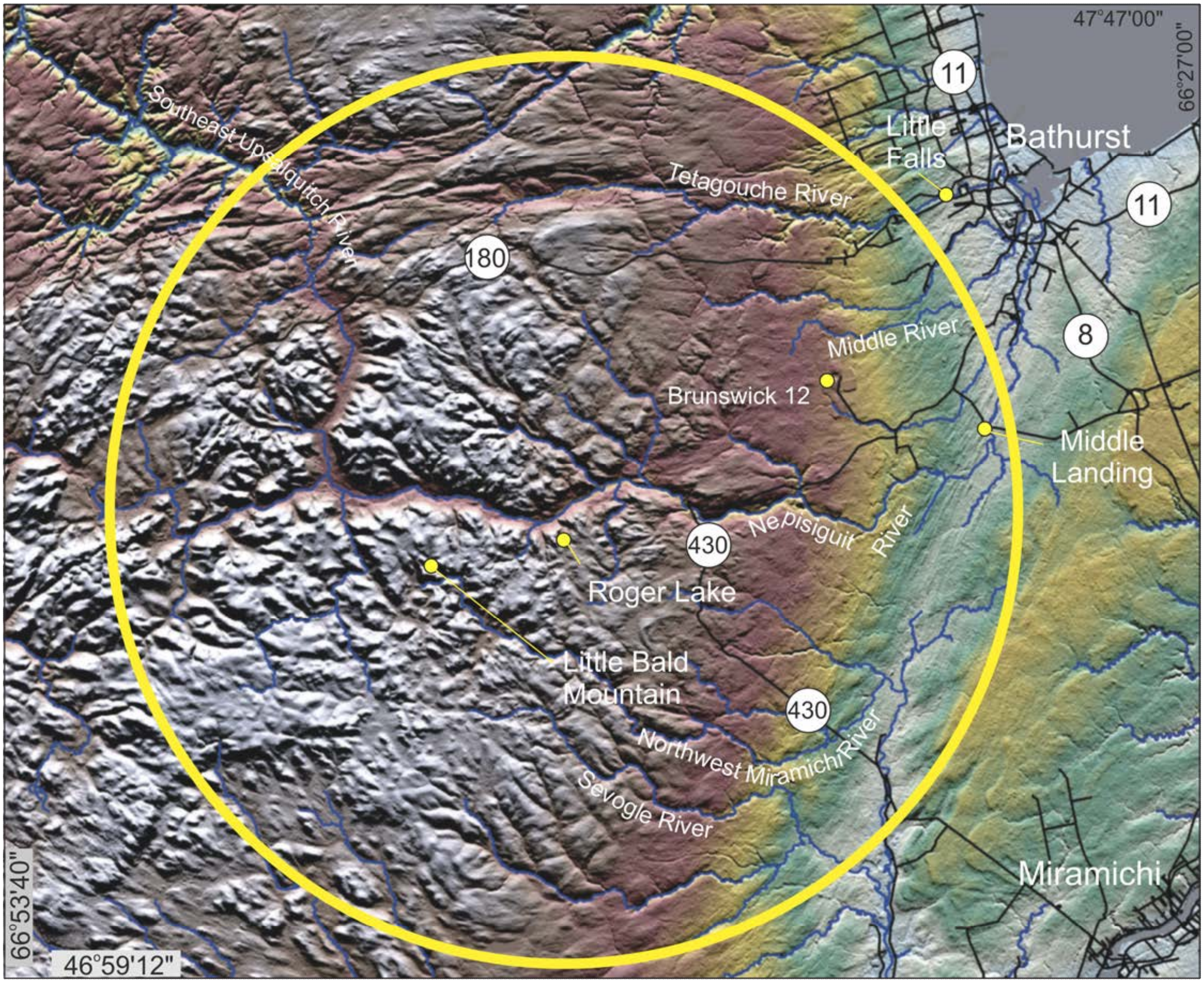

Figure 1. Digital elevation model showing physiographic features and the approximate location of the Bathurst Mining Camp, outlined by the yellow circle. Modified from the 'Satellite Image' of New Brunswick, which was taken from the New Brunswick Department of Energy and Resource Development (DERD) website. Available from https://www2.gnb.ca/content/gnb/en/departments/erd/energy/ content/minerals/content/Surficial_mapping.html.

ly entrenched. The results of this project were published as *Economic Geology Monograph 11* (Goodfellow et al. 2003).

As a result of the EXTECH-II Project, we now know that the BMC comprises five groups of rocks, three of which are approximately coeval and contain Zn-Pb-Cu-Ag VMS deposits (van Staal et al. 2003). We also know that some of these rocks and one VMS deposit (Key Anacon East) extend eastward under Carboniferous cover. However, for the purpose of this paper, the definition of the BMC is restricted to the area that is not concealed by Carboniferous strata, i.e. to the belt of Cambro–Ordovician rocks, which extends southwestward from Bathurst, including the Bathurst Supergroup (predominantly volcanic) and the Miramichi Group (entirely sedimentary).

Camp Overview

The BMC hosts 45 known massive sulphide deposits (Fig. 2), including Brunswick No. 12 (Table 1). Most deposits were discovered in the 1950s and 1960s by airborne geophysical, geological and stream geochemical methods (McCutcheon et al. 2003).

Information Sources

Much has been written about the geology and minerals deposits of the BMC since its discovery in 1952 but the best place to start is with *Economic Geology Monograph 11* (Goodfellow et al. 2003) because the papers contained in this volume provide extensive lists of references to previous work on various topics. There are also two special issues of *Exploration and*

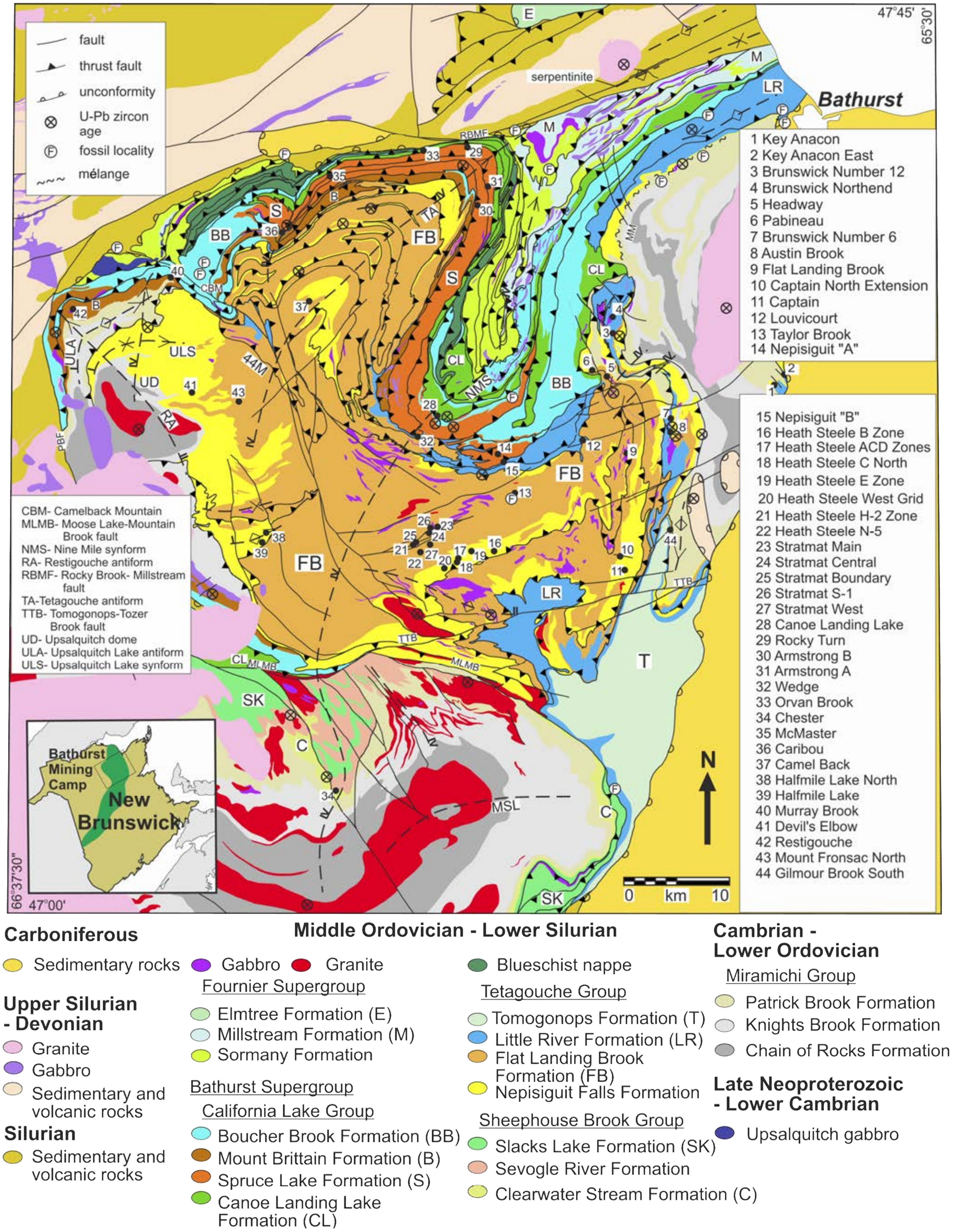

Figure 2. Geological map of the Bathurst Mining Camp showing the distribution of known massive sulphide deposits. Modified from Figure 2 of van Staal et al. (2003).

Table 1. Known massive sulphide deposits in the Bathurst Mining Camp, New Brunswick. Modified from Table 1 of McCutcheon et al. (2003).

								Pb-Zn Resources			Cu				
URN	Deposit	Discoverer	Year	Method	Type	Host	Tonnes	%Pb	%Zn	%Cu	g/t Ag	Tonnes	%Cu	Date	Source
484	Armstrong A	Anaconda	1956	EM	BM	SL	3,377,272	0.42	3.26	0.29	25			1972	Noel 1998
482	Armstrong B	Anaconda	1956	EM	BS	SL						537,400	0.67	1973	Mineral Occurrence Database
069	Austin Brook	Boylen Group	1952	Geol	BM	NF	234,600	3.67	5.68	0.09	82			1991	Luff 1995
054	Brunswick No. 12	Boylen Group	1953	Mag, EM	BM	NF	mined out							2013	
144	Brunswick No. 6	Boylen Group	1952	Mag, EM	BM	NF	858,600	3.01	8.08	0.17	90	1,752,000	1.06	1991	Luff 1995
1263	Brunswick North End	Brunswick Mining & Sm.	1989	Geol	BM	NF	1,011,000	3.00	6.22	0.24	110			1992	Luff 1995
1383	Camel Back	Noranda	1996	AEM, Mag	BM	NF	~200,000	2.7-2.9	3.0-4.0					2006	Walker and Carroll 2006
242	Canoe Landing Lake	Baie Holdings	1960	Soil, EM	SD	CL	1,791,900	0.73	2.67	0.77	44			1995	Johnson 1995
159	Captain	Captain Mines	1956	EM	BS	NF						1,006,000	1.03	2010	Mercator Geol. Services 2011
444	Caribou	Anaconda	1955	Geol, Pros	BM	SL	7,230,000	2.93	6.99	0.43	84			2013	SRK Consulting 2014
071	Chester	Chesterville Mines	1955	AEM	BS	CW						284,000	2,78	2008	Sim and Davis 2008
170	CNE	Sabina	1978	Silt, IP	BM	NF	293,885	2.04	5.62		65			2011	McLaughlin et al. 2011a
285	Devil's Elbow	American Metal Co.	1957	AEM	BS	NF						362,900	1.20	1974	Mineral Occurrence Database
046	Flat Landing Brook	Sabina	1975	AEM	BM	NF	1,700,000	0.94	4.9		20			2006	Walker and Lentz 2006
409	Halfmile Lake	Texas Gulf/Conwest	1955	Geol, EM	BM	NF	5,664,900	2.7	8.27	0.19	34			2010	McLaughlin et al. 2011b
410	Halfmile Lake North	Sweet Grass Oils	1955	EM	BM	NF	597,200	1.4	6.78	0.49	5			2010	McLaughlin et al. 2011b
124	Headway	K. McDonough	1957	Pros	BM	FL	282,572	2.06	6.2	1.43	19			1966	Parliament and Sullivan 1966
395	Heath Steele ACD Zones	American Metal Co.	1954	AEM	BM	NF	553,100	4.18	11.26	0.29	111	113,900	3.56	1998	Heath Steele Mines
396	Heath Steele B Zone	American Metal Co.	1954	AEM	BM	NF	1,439,500	2.38	5.99	1.69	101	597,400	3.18	1998	Heath Steele Mines
395	Heath Steele C-North	Heath Steele Mines	1981	Mag, EM	BM	NF	2,700,000	2.04	6.03	0.39	81			1991	Noranda
397	Heath Steele E Zone	American Metal Co.	1954	AEM	BS	NF	917,000	2.39	5.79	1.47	102			1990	Heath Steele Mines
257	Heath Steele N-5 Zone	Heath Steele Mines	1964	IP	BM	NF	mined out							1991	
1380	Heath Steele West Grid	Heath Steele Mines	1966	IP	BM	NF	961,500	3.12	7.01	0.14	87			1991	Noranda
014	Key Anacon	New Larder U	1953	AM, EM	BM	NF	1,670,000	2.52	6.02	0.14	74			2019	Osisko 2019
1384	Key Anacon East (Titan)	Rio Algom	1993	Geol, EM	BM	NF	290,000	1.57	4.36	0.65	39			2019	Osisko 2019
147	Louvicourt	L. Gray et al.	1964	Pros	BM	FL	136,000	1.23	1.00	0.42	83			1976	Gummer 1976
477	McMaster	Anaconda	1957	EM	BM	SL						250,000	0.75	1972	Mineral Occurrence Database
1418	Mt. Fronsac North	Noranda	1999	Pros, Silt	BM	NF	1,260,000	2.18	7.65	0.14	40			2000	Walker and Graves 2006
414	Murray Brook	Kennco	1956	Silt, EM	BM	MB	5,280,000	1.80	5.24	0.46	69			2017	Puma Exploration 2017
241	Nepisguit "A"	Kennco	1956	EM	BM	SL	1,542,100	0.60	2.80	0.40	10			1976	Mineral Occurrence Database
240	Nepisguit "B"	Kennco	1956	EM	BM	SL	1,360,700	0.40	1.90	0.10	10			1976	Mineral Occurrence Database
239	Nepisiguit "C"	Kennco	1956	EM	BM	SL	635,000	0.70	2.10	0.40	21			1976	Mineral Occurrence Database
062	Orvan Brook	Tetagouche Exp. Co.	1938	Pros	BM	SL	181,000	3.25	6.3		31			1966	Skinner 1974
062	Orvan Brook	Tetagouche Exp. Co.	1938	Pros	BM	SL	uneconomic							1999	Collins 1999
157	Pabineau	Quebec Smelting & Ref.	1953	EM, Geol	BS	NF	136,000	0.87	2.65					1980	Mineral Occurrence Database
139	Restigouche	New Jersey Zinc	1958	Soil	BS	MB	861,882	5.25	7.07	0.33	78			2009	Trevali website 2019
479	Rocky Turn	Anaconda	1957	Pros	BS	SL	131,000	2.69	8.43	0.28	101			1972	Mineral Occurrence Database
255	Stratmat Boundary	Cominco	1961	Soil, EM	BM	FL	mined out							1993	Hamilton and Park 1993
252	Stratmat S0,S1s, S1d, S5	Strategic Minerals	1956	EM	BM	FL	4,700,000	2.10	5.30	0.40	49			2015	Arseneau 2015 (also encompasses URN 1377: Stratmat N-3)
253	Stratmat Central North														
406	Stratmat Main														
xxx	Stratmat New	Noranda	1988	Geol, EM											
256	Stratmat West Stringer	Cominco	1972	IP	BS	FL						181,000	2.00	1981	Mineral Occurrence Database
400	Taylor Brook (Cons. Mor.)	Consolidated Morrison	1977	AEM	BM	FL	179,000	0.64	1.37	0.02	29			2011	McLaughlin et al. 2011a
052	Wedge	Cominco	1956	Geol, Pros	BM	SL	545,000	1.71	5.21	1.75				1992	SLAM Exploration 2017
							48,520,711					**5,084,600**			

URN = Unique Record Number; Method: AM = airborne magnetic; AEM = airborne electromagnetic; EM = electromagnetic; Geol = geology; IP = induced polarization; Mag = magnetic; Pros = prospecting; Silt = silt geochem; Soil = soil geochem; Type: BM = stratiform bimodal volcanic or sediment-hosted massive sulfides; BS = stratabound bimodal volcanic or sediment-hosted disseminated and stringer sulfides; SD = stratiform sediment-hosted sulfides; Host: CL = Canoe Landing Lake Fm; CW = Clearwater Stream Fm; FL = Flat Landing Brook Fm; MB = Mount Brittain Fm; NF = Nepisiguit Falls Fm; SL = Spruce Lake Fm; Fm =Formation; Date = Year calculation was done; Note: Compiled by W.M. Luff (May, 1999); does not include Production; 5 deposits do not have calculated estimates but all are < 1 million tonnes for a total of about 2 million tonnes.

Mining Geology that are devoted to mineral deposits of the BMC; one is Volume 1, No. 2 (Davies et al. 1992), and the other is Volume 15, No. 3-4 (Lentz 2006). Finally, published maps and reports of the Geological Survey of Canada and the Geological Surveys Branch of New Brunswick Department of Energy and Resource Development are too numerous to list but can be found on the websites of these two organizations: https://www.nrcan.gc.ca/earth-sciences and http://www2.gnb.ca/content/gnb/en/departments/erd/energy/content/minerals.html, respectively. On the New Brunswick website, on-line databases also contain a wealth of information, including a "*Bedrock Lexicon*" that has a description of every formally named rock unit (formations, groups and plutons) in the province.

Historical information about exploration in the BMC comes from published books and articles. Some notable books are: *Metals and Men* (LeBourdais 1957), which contains a few pages on the discovery of the first deposits; *The Discoverers* (Hanula 1982), which contains a section on "The Bathurst-Newcastle Area"; *The Birth of the Bathurst Mining Camp* (Belland 1992), which describes the development history of the Austin Brook Iron Mine and Brunswick No. 6 base metal deposit; and *Gesner's Dream* (Martin 2003), which contains three chapters relating to the history of mineral exploration in the Bathurst area, prior to and leading up to the discovery of Brunswick No. 6. In addition, articles in *The Northern Miner*, and New Brunswick newspapers, such as *The Northern Light*, *The Daily Gleaner*, and *Telegraph Journal* are too numerous to list. Finally, there is unpublished correspondence in the files of NB Department of Energy and Resource Development, which provides useful historical information.

GEOLOGICAL SETTING

Regional Geology

The Bathurst Mining Camp is situated towards the northern end of a northeasterly trending belt or terrane of Cambrian to Ordovician, sedimentary and volcanic rocks (Fig. 2, shaded area on inset). This terrane is unconformably overlain by or in fault contact with Silurian rocks to the north and west, and unconformably overlain by Carboniferous rocks to the east. The BMC contains rocks that belong to the Dunnage and Gander zones of the Canadian Appalachians (cf. Williams 1979) and formed during the Ordovician by rifting of an existing submarine volcanic arc (Popelogan Arc) on the continental margin of Ganderia (Fig. 3a), what was then the eastern margin of the Iapetus Ocean. Rifting ultimately produced a Sea of Japan-style basin that is referred to as the Tetagouche-Exploits back-arc basin (van Staal et al. 2003).

This Japan-style basin opened by rifting of continental crust in the Early Ordovician, initially giving rise to felsic volcanism (Figs. 3b, c) followed by mafic volcanism (Fig. 3d) and opening of a back-arc rift basin. It ultimately closed by northwest-directed subduction during the Late Ordovician to Early Silurian (van Staal et al. 2003). Dunnage Zone rocks are represented by the approximately coeval California Lake, Sheephouse Brook, and Tetagouche groups, which disconformably overlie sedimentary rocks of the Miramichi Group (Gander Zone). They are structurally overlain by oceanic crustal rocks of the Sormany Group (formerly included in the Fournier Group). Each of the first three groups is characterized by variable proportions of felsic and mafic volcanic rocks, which were deposited in different parts of the basin but were later tectonically juxtaposed in a Subduction–Obduction accretionary wedge, i.e. the Brunswick Subduction Complex (van Staal 1994). The bulk compositions of volcanic rocks and U–Pb radiometric dating show that each of these groups evolved from felsic- to mafic-dominated volcanism through time. This change in volcanism is interpreted to reflect crustal thinning during the rifting process, i.e. extended continental to transitional oceanic crust, respectively. Most of the VMS deposits are associated with the early-erupted felsic volcanism in each of these groups.

Camp Geology

Glacial History

The glacial history of the BMC is described in detail by Parkhill and Doiron (2003), which includes a reference list of prior work on the Quaternary geology of the area. The following description is extracted from their paper.

The entire BMC was covered by ice during the last glacial maximum, referred to as the Wisconsinan (75,000 to 10,000 BP). Ice cover is indicated by the presence of a single homogeneous basal till over most of the area, erratics (exotic pebbles and boulders) in till at the highest elevations, and a similar glacial history in adjacent areas of northern New Brunswick. The preservation locally of pre-glacially weathered bedrock (grus) indicates that glacial erosion in some parts of the BMC was relatively weak, generally on the down-ice sides of hills.

The BMC was affected by multiple phases of ice flow (Parkhill and Doiron 2003), as indicated by the orientations of both small-scale features (striations, grooves, roches moutonnées, glacially sheared bedrock, and till fabrics) and large-scale features (fluted bedrock, eskers, and DeGeer moraines). A main eastward ice movement (California and Jacquet flow patterns) was followed by northeastward flowing ice (Belledune flow pattern) in the northern and central parts of the BMC. In the southern BMC, ice flow was mainly toward the southeast (Sevogle flow pattern) but in the eastern and southeastern parts, the dominant ice movement was north-northeastward (Nepisiguit flow pattern). These trends can be seen on the satellite image (Fig.1). Understanding the variable ice flow directions is important for interpreting till geochemical anomalies.

In the BMC, till thickness is variable (0–5 m) and till is locally derived (< 1 km transport). A strong correlation exists between the lithology of clasts/pebbles in till and underlying bedrock; commonly, more than 75% of them are derived from the directly underlying rock unit. However, some clasts and boulders (erratics) can be found up to 20 km down-ice from their source. Glacial dispersal patterns of clasts generally indicate east to northeast ice-flow directions. In areas where till thickness is >5 m, clasts of directly underlying bedrock are less

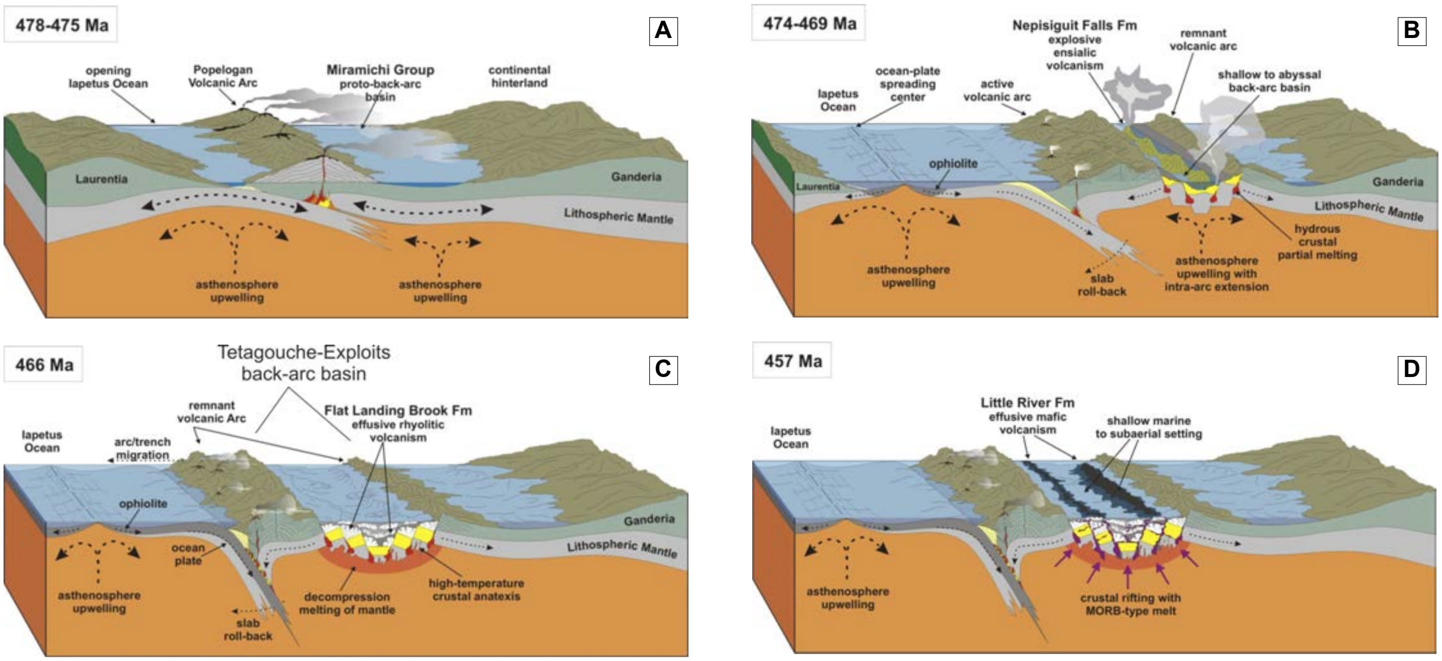

Figure 3. Schematic block diagrams portraying four time intervals: a) 478–475 Ma, b) 474–469 Ma, c) 466 Ma, and d) 457 Ma, and the depositional settings of various Ordovician rock units of the Tetagouche Group. From Figure 1.4 of Wills (2014).

numerous or even absent, and there is no geochemical expression, at surface, of underlying base metal deposits, as for example at the CNE deposit (Table 1).

Bedrock Geology

Since descriptions of all the rock units in the BMC (Fig. 2) are available on-line in the *Bedrock Lexicon* [http://dnr-mrn.gnb.ca/Lexicon/Lexicon/Lexicon_Search.aspx?lang=e], the focus here is on those units that host VMS deposits. They are the California Lake, Sheephouse Brook, and Tetagouche groups, which are shown schematically in Figure 4 with their constituent formations. Each is described below.

The California Lake Group comprises the Mount Brittain, Spruce Lake, Canoe Landing Lake and Boucher Brook formations. The latter formation overlies each of the first three in separate nappes that are named after their characteristic volcanic unit (van Staal et al. 2003). The Mount Brittain nappe is the structurally lowest and the Canoe Landing Lake nappe is the highest. Collectively, the three nappes make up the California Lake Group and account for about 15% of the surface area of the BMC (Fig. 2). The Mount Brittain, Spruce Lake and Canoe Landing Lake formations are coeval, and all contain massive sulphide deposits.

Mount Brittain Formation: This formation hosts 2 of the 13 deposits in the California Lake Group (Table 1). It is predominantly composed of feldspar-crystal and lithic felsic tuff with minor aphyric felsic rocks, but it also includes a thin sedimentary unit (Charlotte Brook Member) at the base, which gradationally overlies rocks of the Miramichi Group. This sedimentary unit, which hosts the Murray Brook deposit, comprises dark grey shale and wacke with a few thin tuff beds. Even though this formation is about the same age as the Spruce Lake Formation (Fig. 4), no direct linkage (inter-fingering) between the two formations exists, all mutual contacts being tectonic. Lithologically, the Mount Brittain feldspar-crystal tuff more closely resembles quartz-poor crystal tuff of the Nepisiguit Falls Formation than it does crystal tuff of the Spruce Lake Formation.

Spruce Lake Formation: This formation hosts 10 of the 13 deposits in the California Lake Group (Table 1). It mainly comprises feldspar-phyric to aphyric felsic volcanic rocks, in places with intercalated basalt. Some dark grey to black, fine-grained sedimentary rocks, which overlie, underlie and/or are interbedded with this volcanic pile, are also included in this formation. The basalt is correlative with rocks in the Canoe Landing Lake Formation (van Staal et al. 2003) and shows that there was a spatial association between these two formations.

Canoe Landing Lake Formation: This formation hosts only 1 of the 13 deposits in the California Lake Group (Table 1). It predominantly consists of pillow basalts and associated rocks, including interflow chert and red shale, but also contains some fine-grained, dark grey sedimentary rocks and minor felsic volcanic rocks. The felsic volcanic rocks are lithologically like those in the Spruce Lake Formation.

The Sheephouse Brook Group comprises the Clearwater Stream, Sevogle River and Slacks Lake formations. The first formation (Fig. 4) hosts the only known deposit (Chester) in the group (Table 1). This group makes up about 5% of the surface area of the BMC (Fig. 2).

Clearwater Stream Formation: Fyffe (1995) defined this formation, which hosts the Chester deposit, as "*the plagioclase-phyric felsic volcanic rocks that immediately overlie sedimentary rocks of*

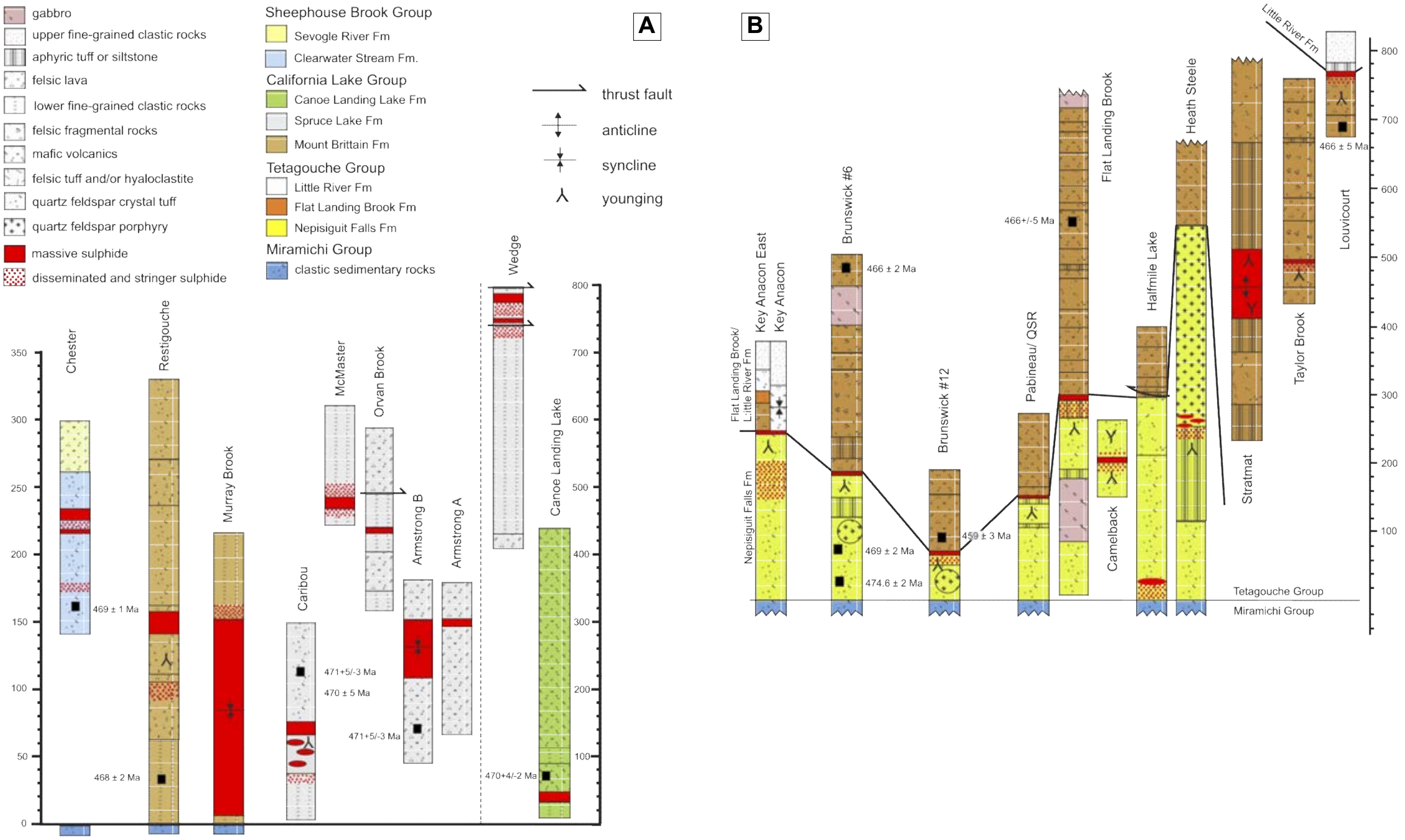

Figure 4. Schematic stratigraphic columns from deposits in various parts of the Bathurst Mining Camp: a) Selected drill-hole columns from deposits in the California Lake Group and one (Chester) from the Sheephouse Brook Group, b) Selected drill-hole columns from deposits in the Tetagouche Group. Modified from figures 7 and 10 of McCutcheon et al. (2001). QSR = Quebec Sturgeon River.

the Patrick Brook Formation south of the Moose Lake shear zone." These volcanic rocks are approximately coeval (Fig. 4) with the felsic volcanic rocks in the California Lake Group and the Nepisiguit Falls Formation (Tetagouche Group; see below).

The Tetagouche Group comprises the Nepisiguit Falls, Flat Landing Brook, and Little River formations, in ascending stratigraphic order. The Tetagouche Group occurs in two major nappes that constitute approximately half of the surface area of the BMC (Fig. 2). Both the Nepisiguit Falls and Flat Landing Brook formations contain massive sulphide deposits.

Nepisiguit Falls Formation: This formation hosts 24 of the 32 deposits in the Tetagouche Group, and the bulk of the massive sulphide tonnage in the BMC (Table 1). The age of the Nepisiguit Falls Formation is constrained by several U–Pb isotopic ages, which suggest an age of circa 470 Ma (Fig. 4). Rocks of this formation were commonly referred to as 'quartz augen (eye) schists' (QAS or QES) and 'quartz-feldspar augen schists' (QFAS) in the pre-1990 literature (e.g. Skinner 1974, p. 29–30). The former rock type is merely an altered version of the latter.

At the type locality on Nepisiguit River, this formation is divisible into two parts. The lower third comprises massive quartz-feldspar 'porphyry,' and the upper two thirds comprises medium- to coarse-grained, quartz–feldspar-rich, volcaniclastic rocks that are interlayered with aphyric tuff. The quartz-feldspar 'porphyry' conformably overlies sedimentary rocks of the Miramichi Group. It typically has a vitreous, cryptocrystalline groundmass, contains less than 30% crystals (up to 15 mm), and lacks any evidence of reworking. The volcaniclastic rocks (crystal tuff) appear to conformably overlie the massive quartz-feldspar 'porphyry' and generally become finer grained and thinner bedded up section (McCutcheon et al. 1993a, 1997). They also contain abundant (30% or more), commonly broken and rounded, quartz and feldspar (mostly < 5 mm) in a very fine-grained granular matrix. They exhibit primary features such as crystal sorting, graded beds and rare pseudomorphed pumice clasts. In the upper part of the section, some beds contain rare, lapilli-sized, lithic clasts of aphyric (ash) tuff or rhyolite and, at the top of the section, chloritic mudstone and silicate iron formation are present.

In the Brunswick Belt (between Brunswick No. 6 and 12), the number of eruptive/emplacement units in the Nepisiguit Falls Formation ranges from 2 to 6 (McCutcheon and Walker 2007) and individual units range from a few metres to a few tens of metres in thickness and are locally separated by narrow intervals of fine-grained sedimentary material. At the top of the Nepisiguit Falls Formation, chloritic and locally magnetic mudstone (silicate iron formation) is interbedded with dark greenish grey, fine-grained volcaniclastic rocks, which constitute the "Brunswick Horizon" at the nearby Austin Brook and Brunswick No. 6 mine sites (Fig. 2). This Algoma-type Fe-Formation and the massive sulphide deposits along the Brunswick Belt are collectively referred to as the Austin Brook Member. The contact with massive rhyolite of the overlying Flat Landing Brook Formation appears to be conformable.

Elsewhere, the Nepisiguit Falls volcanic pile exhibits lateral variations in thickness and proportions of rock types. At Little Falls on Tetagouche River (Fig. 1), the section is approximately 30 m thick and mainly composed of interbedded aphyric tuff and fine-grained crystal tuff, with isolated lenses (channels) of coarse-grained volcaniclastic rocks. The coarse-grained rocks contain more than 50% crystals (quartz and feldspar) and a few intraformational clasts. Quartz-feldspar 'porphyry' is conspicuously absent. This section overlies calcareous rocks of the Vallee Lourdes Member (formerly formation), which unconformably overlies the Patrick Brook Formation of the Miramichi Group. At Heath Steele (Fig. 2), the Nepisiguit Falls Formation contains 'porphyry', but it overlies volcaniclastic rocks rather than underlies them as in the type section. The volcaniclastic rocks are interbedded with quartz wacke and carbonaceous shale, which are typical of the Patrick Brook Formation (Lentz and Wilson 1997). This implies that the contact between the Tetagouche and Miramichi groups is conformable at this locality rather than disconformable as it is in some places (cf. van Staal 1994).

Flat Landing Brook Formation: This formation hosts 8 of the 31 deposits in the Tetagouche Group (Table 1). It comprises aphyric to feldspar-phyric (± quartz) felsic flows, hyaloclastite, and crackle breccia, interbedded with minor aphyric tuff, basalt, mudstone and iron formation (silicate magnetite and Fe-Mn types). Feldspar ± quartz phenocrysts are small (1–3 mm) and constitute less than 10% of the rocks; the groundmass is cryptocrystalline. Aphyric tuff and basalt appear to be most abundant in the northwestern (upper) part of the BMC where they constitute separate mappable members. The Flat Landing Brook Formation is a few million years younger than the Nepisiguit Falls Formation (Fig. 4).

Structure

Rocks of the BMC have undergone four phases of deformation in the Late Ordovician to Early Silurian, as a result of amalgamation in an accretionary wedge environment known as the Brunswick Subduction Complex (van Staal 1994). They are variably deformed and metamorphosed to greenschist facies, including local blueschist that reflects burial depths of 11 km or more in a subduction zone. Two cleavages are common in the rocks; however, four can be discerned in places, and conversely, none is visible locally.

Mineralization

Many of the massive sulphide deposits of the BMC were originally deposited as sulphide mounds in a relatively deep ocean basin at or near the sea floor from so-called 'black smokers', which are nothing more than metal-rich, magmatically-heated, hot springs. To be called a massive sulphide deposit, sulphide minerals must constitute more than 60% of the rock (Franklin et al. 2005). The predominant sulphide minerals in most BMC deposits are pyrite, pyrrhotite, sphalerite, galena, and chalcopyrite (Goodfellow and McCutcheon 2003); the galena tends to be silver-rich. Other sulphide minerals occur in minor amounts, including arsenopyrite, marcasite, and stannite; some sulphosalt and oxide minerals are also present. The relative proportions of sulphide minerals, which are very fine-grained and commonly layered, vary depending upon the primary depositional facies of each deposit and the amount of tectonic (post-depositional) recrystallization and mobilization that has occurred. Many of the deposits have an oxide–silicate–carbonate 'iron formation' that caps and extends laterally beyond the limits of the massive sulphide facies. Notably, the Fe_{Total}/Mn, Fe/Ti, Ba/Ti, Eu/Eu*and Pb/Zn ratios (and several other element ratios) in these iron formations tend to increase toward the massive sulphide facies (Peter and Goodfellow 1996, 2003).

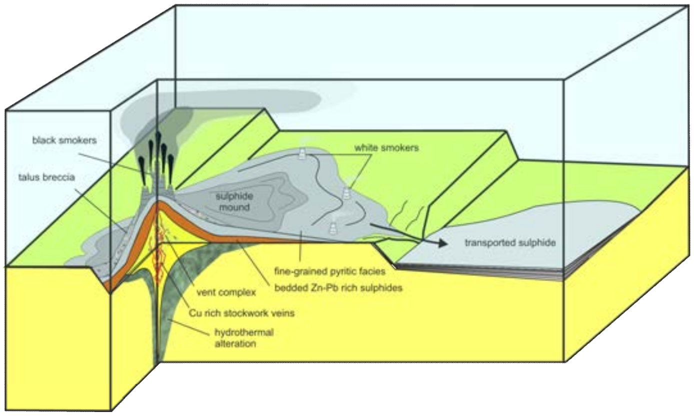

Figure 5. Idealized block model showing a sulphide mound on the sea floor. From Figure 1 of Tornos et al. (2015).

An idealized block model showing a sulphide mound on the sea floor is shown in Figure 5. The mound comprises a debris field of collapsed sulphide chimneys built over an alteration pipe that represents the pathway (plumbing) from depth to surface of hot (hydrothermal), metal-bearing fluids. High-temperature, copper-bearing sulphide minerals and pyrrhotite tend to be deposited in the throat of the alteration pipe, i.e. the vent complex, whereas lower temperature lead and zinc sulphide minerals and pyrite are deposited in the mound. As the mound grows over time, early-formed low temperature phases in the lower part of the mound are dissolved and re-precipitated in the upper more distal parts of the mound in a process that is called 'zone refining' (Ohmoto 1996).

Other massive sulphide deposits in the BMC did not form on the sea floor but were deposited from hydrothermal fluids beneath the sea floor, which reacted and replaced volcanic glass and other soluble components in permeable layers. Such deposits have similar mineralogy as the ones deposited on the sea floor, but they are generally not layered, and do not have a capping iron formation.

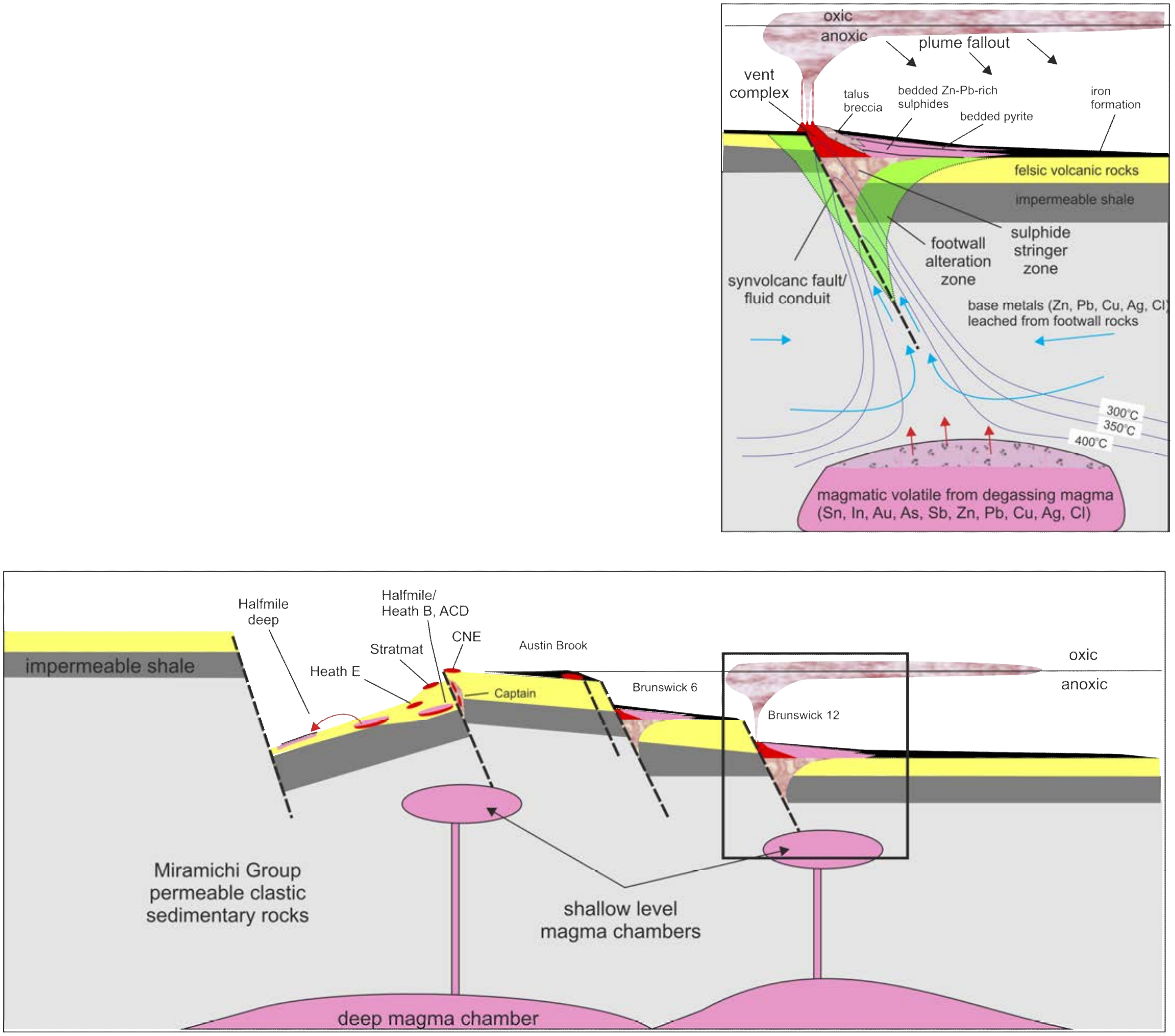

Figure 6. An idealized cross section showing the conceptual ore controls of Brunswick-type deposits. Modified from Figure 28 of Goodfellow and McCutcheon (2003). ACD = A, C, and D zones; CNE = Captain North Extension.

Deposit Characteristics

An idealized cross section showing the conceptual ore controls of Brunswick-type (Tetagouche Group) deposits is shown in Figure 6. Initially, a deep, large-volume, felsic magma chamber supplied felsic volcanic rocks (Nepisiguit Falls Formation) to the sea floor; then as pyroclastic volcanism waned, high-level (sub-volcanic), small-volume magma chambers formed and cooled *in situ* (Fig. 6). These small-volume magma bodies were probably localized along the ring-fracture zone of the caldera that must have been created by the initial pyroclastic volcanism. Each small-volume magma body supplied heat and hot metal-bearing fluids that were channeled to surface via a synvolcanic fault (feeder zone) and created a hydrothermal system (black smoker) that formed a sulphide mound on the sea floor. These magma bodies also supplied late-stage, coarse-grained, quartz-feldspar porphyry sills that intruded the felsic volcanic pile. The hydrothermal fluids altered the wall rocks adjacent to the feeder zone, creating a range of alteration minerals that reflect the temperature of the fluid and the original composition of the wall rocks.

The alteration is zoned within and away from the feeder zone, and the mineralization is also zoned (Goodfellow and McCutcheon 2003). The most intense alteration (Zone 1) is characterized by silicification, iron-chlorite and disseminated sulphide minerals, and is at the top of the feeder zone directly beneath massive sulphide of the 'vent complex'. Zone 2 alteration is deeper in the feeder zone and characterized by iron-chlorite, some sericite, and abundant sulphide stringers that disappear with increasing depth in the pile. Zone 3 alteration is outboard of the feeder zone, is feldspar-destructive (resulting in 'quartz-augen schist') and is characterized by iron-magnesium chlorite and sericite. Zone 4 is outboard of Zone 3, only partially feldspar destructive, and is characterized by phengite and magnesium chlorite. The stringer sulphide minerals in the 'sulphide-stringer zone' predominantly comprise pyrrhotite and/or pyrite with chalcopyrite and traces of sphalerite and galena. The 'vent complex' facies comprises pyrrhotite and/or pyrite breccia that is replaced and/or veined by pyrrhotite, pyrite, chalcopyrite, magnetite, chlorite, quartz, and siderite in variable proportions. The 'bedded pyrite' facies mainly consists of fine-grained, massive pyrite with minor sphalerite, galena, and chalcopyrite. The 'bedded ore' facies generally comprises interlayered, fine-grained pyrite, brown sphalerite, and galena with minor amounts of arsenopyrite, marcasite, cassiterite, stannite, tetrahedrite, and bournonite.

To summarize, the primary characteristics of Brunswick-type mineralization are: 1) stratigraphic position at, or near the top of, the Nepisiguit Falls Formation; 2) associated quartz-feldspar porphyry sills; 3) an underlying feeder zone characterized by stringer sulphide minerals (pyrrhotite and chalcopyrite); 4) proximal iron-chlorite alteration and silicification; 5) an outboard, iron-magnesium chlorite and sericite alteration that is feldspar-destructive; 6) various sulphide facies (vent complex, bedded pyrite, and bedded ore) that appear to reflect deposition in different parts of an original sulphide mound; and 7) an oxide-silicate-carbonate iron formation that caps and extends laterally beyond the massive sulphide.

Primary characteristics for massive sulphide deposits in other parts of the BMC show similarities and differences to those in the Tetagouche Group (Brunswick type). For example, other deposits are also spatially associated with felsic volcanic rocks but not necessarily hosted by them; rather, they are in sedimentary rocks that either overlie or underlie the deposits (e.g. Canoe Landing Lake, Caribou, and Murray Brook). Quartz-feldspar porphyry sills are absent. Some deposits are within felsic volcanic rocks but largely formed beneath the sea floor by replacement of permeable breccia and/or hyaloclastite (e.g. Restigouche, Taylor Brook). Oxide iron formation is absent in deposits of the California Lake and Sheephouse Brook groups. The main sulphide minerals are the same, i.e. pyrite, pyrrhotite, sphalerite, galena, and chalcopyrite, but galena tends to be less silver rich. Also, deposits in the California Lake Group tend to have more trace gold than those in the Tetagouche Group. The associated hydrothermal alteration is similar, i.e. proximal iron-chlorite alteration with outboard, iron-magnesium chlorite and sericite alteration. Most deposits do not have a well-defined stringer zone (probably tectonically cut out), with the exception of the Chester deposit in the Sheephouse Brook Group.

Secondary ore controls on mineralization of all types exist in the BMC. Tectonic thickening of sulphide units occurs in F_1–F_2 fold hinges with thinning on fold limbs (van Staal and Williams 1984). Tectonic remobilization of ductile sulphide occurs locally, e.g. the pyrrhotite breccia units at Heath Steele (de Roo et al. 1991). Recrystallization of sulphide minerals in the contact aureoles of younger felsic intrusions increases their grain size, e.g. Key Anacon, Chester.

HISTORY OF EXPLORATION AND DISCOVERY

The history of mineral exploration in the Bathurst Mining Camp (BMC) can be divided into six periods: a) pre-1952, b) 1952–1958, c) 1959–1973, d) 1974–1988, e) 1989–2000, and f) post-2000. The history of mine development and mineral production follows a different time-line and is described in a companion paper (*The Bathurst Mining Camp Part 2: Mining History and Contributions to Society*). Much of the following description is extracted from McCutcheon et al. (2003); however, the pre-1952 events are more thoroughly described by Martin (2003).

Pre-1952: Prior to Discovery

Prior to the discovery of Brunswick No. 6 in 1952, the 'Bathurst District' was known for its 'Nipisiguit (sic) Iron Ore Deposit', now called Austin Brook (Table 1, Fig. 2). The history of discovery and development of this deposit is thoroughly described by Belland (1992) and Martin (2003).

In 1938, the Orvan Brook massive sulphide deposit (Table 1, Fig. 2) was found by a prospector from Nevada, Mr. Dan Sheahan, who was working for the Tetagouche Exploration Company (Wright 1939; Martin 2003). The company drilled 28 holes, outlining a deposit at least 1900 m in strike length, and demonstrated for the first time that massive sulphide occurs in the Bathurst area (Tupper 1969).

In the same year that Orvan Brook was discovered, T. LaFrance of Bathurst "*opened some promising looking leads at Middle Landing on the Nepisiquit (sic) river*" (Wright 1939). However, it was not until 1946 that P.J. Leger of Bathurst acquired the mineral rights to this prospect (Wright 1947). Subsequently, a 14-hole diamond-drilling program intersected copper-bearing, vein-sulphide, but massive sulphide deposits were not found (unpublished company report by M.A. Cooper 1947). We now know that the Leger Cu prospect is part of the nearby Key Anacon massive sulphide deposit (Table 1; Fig. 2).

Mr. Patrick (Paddy) W. Meahan (Fig. 7a), a mining engineer, was the catalyst that caused attention to be focused on the mineral potential of the Bathurst area in the late 1940s (McCutcheon et al. 1993b), but Dr. William J. Wright and Dr. Graham S. MacKenzie (Fig. 7b, c) set the stage (Martin 2003). Dr. Wright, the first Provincial Geologist, and Dr. MacKenzie, were both teaching at the University of New Brunswick, and working collaboratively to promote the mineral potential of New Brunswick (Martin 2003). In 1943, Dr. MacKenzie was contracted by Wright to prepare a plan and report on the geology of the iron mine at Austin Brook, which Dominion Steel and Coal Corporation Ltd. had acquired the previous year (Martin 2003). The samples that he collected at Austin Brook played an important role in the lead-up to the Brunswick No. 6 discovery (McCutcheon et al. 1993a; Martin 2003).

In 1951, A. B. Baldwin (Fig. 7d), a graduate student of Dr. MacKenzie, was working on his master's thesis, with the aid of a $3000 James Dunn Scholarship (Belland 1992). The subject of his thesis was the Hayot Lake iron formation in Labrador,

Figure 7. Photos: a) Mr. Patrick (Paddy) W. Meahan c. 1965, b) Dr. William J. Wright c. 1930, c) Dr. Graham S. MacKenzie c. 1940, d) Mr. A. Bennet (Ben) Baldwin 1951. All photographs from the New Brunswick Department of Energy and Resource Development archives.

Figure 8. Photos: a) Mr. M. James (Jim) Boylen c. 1960, b) Mr. Robert (Bob) J. Issacs c. 1970, c) Dr. C. Phillip (Phil) Jenney 1955, d) Dr. Walter Holyk 1955. The first two photographs are from the Canadian Mining Hall of Fame; the last two are from the New Brunswick Department of Energy and Resource Development archives.

but one of the conditions of his scholarship was that the thesis should be, at least in part, about New Brunswick. Dr. MacKenzie suggested that he could satisfy this requirement by comparing the Hayot Lake iron formation with samples from Austin Brook, which he had collected in 1943. In examining polished sections of those samples, Baldwin observed base-metal sulphide minerals in the footwall pyrite zone of the Austin Brook deposit.

In 1952, the Austin Brook property was under license to Brudon Enterprises Limited of Montreal; Baldwin's findings were communicated to the company along with a recommendation to explore the property for base metals (MacKenzie 1958). Brudon did not follow up on MacKenzie's recommendation but instead chose to offer the Austin Brook concession to M. J. (Jim) Boylen (Fig. 8a), a prominent Toronto mining man (Martin 2003). Boylen was representing a small group of New York investors, who that year had put $1 million in the 'M.J. Boylen Nominee Account' to find a mine in Canada (*The Northern Miner*, January 15, 1953).

In the spring of 1952, Meahan was acting as an independent scout for Boylen, who earlier that year had purchased Meahan's Elmtree Pb–Zn property, which would become the Keymet Mine (McCutcheon et al. 1993a; Martin 2003). Meahan learned of Baldwin's findings and independently sampled and assayed the footwall pyrite zone at Austin Brook (Martin 2003). One sample returned values of 9% Zn and 4% Pb (Hanula 1982), so he advised Boylen to acquire the Austin Brook property from Brudon. Consequently, Boylen optioned the Austin Brook property in the summer of 1952 (Belland 1992), which was the sixth project financed by the 'M.J. Boylen Nominee Account'. Robert (Bob) J. Issacs (Fig. 8b), Boylen's chief mining engineer, insisted that a ground electromagnetic (EM) survey be conducted to guide drilling on the property

(Martin 2003). Diamond drilling at Austin Brook began in late August; results from the first 11 holes were negative but Hole B-12, located on an EM anomaly approximately 1000 m north of Austin Brook, intersected the Brunswick No. 6 deposit on October 22nd (Belland 1992). The Bathurst Mining Camp was born.

1952–1958: The Discovery Years

Hole B-12 intersected approximately 100 m of massive sulphide and sparked a chain of events, which were described by Belland (1992): 1) Issacs told Meahan to secure the drill core and say absolutely nothing to anyone about what had been found; 2) Boylen formed Brunswick Mining and Smelting Corporation Limited (and two other companies, Martin 2003) on October 31, 1952, with head office in Saint John; 3) Fifteen additional holes were drilled in the No. 6 deposit; 4) Approximately 1000 mineral claims were staked (*The Northern Miner*, January 15, 1953) by Boylen interests, over magnetic anomalies north and south of the No. 6 Project; 5) The drill core was sent for assay to the geochemistry laboratory at St. Francis Xavier University, Nova Scotia, rather than a commercial laboratory in Ontario, to help keep the discovery a secret; 6) The discovery was announced on the front page of *The Northern Miner* on January 15, 1953.

A staking rush was predicted in the January 15, 1953 issue of *The Northern Miner* and it came to pass a week later. The headline in the January 29 edition read "*Frenzied staking in N.B. spreads far and wide*". The number of mineral claims in the Province went from a few thousand in effect at the beginning of the year to over 41,000 claims (approximately 665,500 ha) by the end of the year. At the peak in 1956, over 43,000 claims (approximately 825,200 ha) were in effect; at no time since has there been more land claimed in New Brunswick, even during the uranium staking rushes of the early 1980s and 2007-08, or the flow-through gold rush of the late 1980s (Fig. 9). From 1953 to 1957, *The Northern Miner* had numerous articles about mining properties and activity in the Bathurst–Newcastle area.

The discoveries of the 1950s largely resulted from the application of geophysical methods (Seigel 1956; Ward 1958). Initially, ground exploration, using electromagnetic (EM) methods, focused on areas with airborne magnetic (Mag) anomalies. In early 1953, Boylen staked the second largest magnetic anomaly known in the area, located 9.7 km north of Brunswick No. 6. (Skinner 1974). A ground electromagnetic survey carried out that same year outlined a strong anomaly about 610 m east of the crest of the aeromagnetic anomaly (Skinner 1974), and subsequent drilling revealed the Anacon-Leadridge (renamed Brunswick No. 12) orebody (Table 1; Fig. 2). Drilling of another ground EM anomaly in 1953 resulted in the discovery of the New Larder 'U' (renamed Key Anacon) deposit. In 1954, Little River (renamed Heath Steele) was discovered as a result of an airborne electromagnetic (AEM) survey that was conducted by the American Metal Company (Jenny (sic) 1957), the very first in the world.

Dr. C.P. (Phil) Jenney (Fig. 8c), Canadian exploration manager for the American Metal Company (AMCO), had negotiated an agreement with the International Nickel Company (INCO) to fly an AEM survey in New Brunswick in 1953 (Gallagher 1999). Numerous AEM anomalies were found and the second hole of a follow-up drilling program intersected the A-Zone in 1954. Drilling of other AEM anomalies led to the discovery of the B, C, D, and E zones.

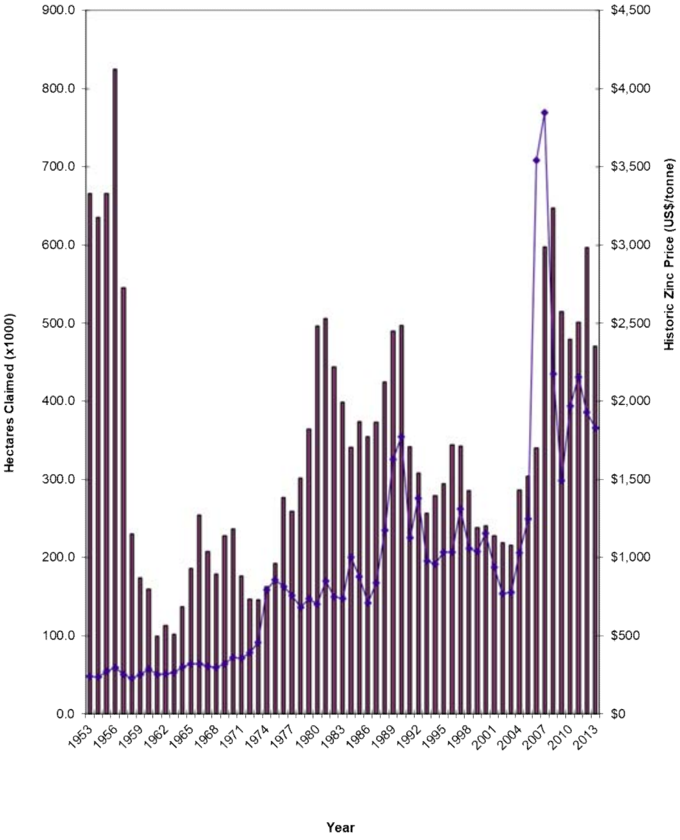

Figure 9. Bar graph showing numbers of hectares claimed in New Brunswick from 1953 to 2013, with the historic price of zinc (in US$/tonne) superimposed. Modified from Figure 2 of McCutcheon et al. (2003).

After the discovery of the Brunswick orebodies, air photos became widely used in conjunction with aeromagnetic maps as a means of selecting properties for exploration, e.g. Halfmile Lake. Dr. Walter Holyk (Fig. 8d), geologist with the Middle River Mining Company Limited (Texasgulf Sulphur Corporation), staked a block of thirty claims in the Halfmile Lake area based on data from aeromagnetic maps, aerial photographs and reconnaissance geology (Holyk 1957 and personal communication 2000). Airborne electromagnetic and magnetic surveys were performed over these claims, followed by a ground EM survey (Holyk, in Hanula 1982). The first hole drilled on the property in late 1955 discovered the Halfmile Lake deposit.

In the latter part of 1954, "The Anaconda Company (Canada) Limited" contracted Dr. Cameron G. Cheriton (Fig. 10a) to do an initial geological study of the Bathurst area (Cheriton 1960). He focused his efforts on the northern part of the BMC where Holyk was not exploring and determined that the New Calumet Zone (Orvan Brook) is hosted by intravolcanic sedimentary rocks. He traced these rocks east and west, using aeromagnetic maps, aerial photographs and ground traverses, to delineate favorable areas for further exploration.

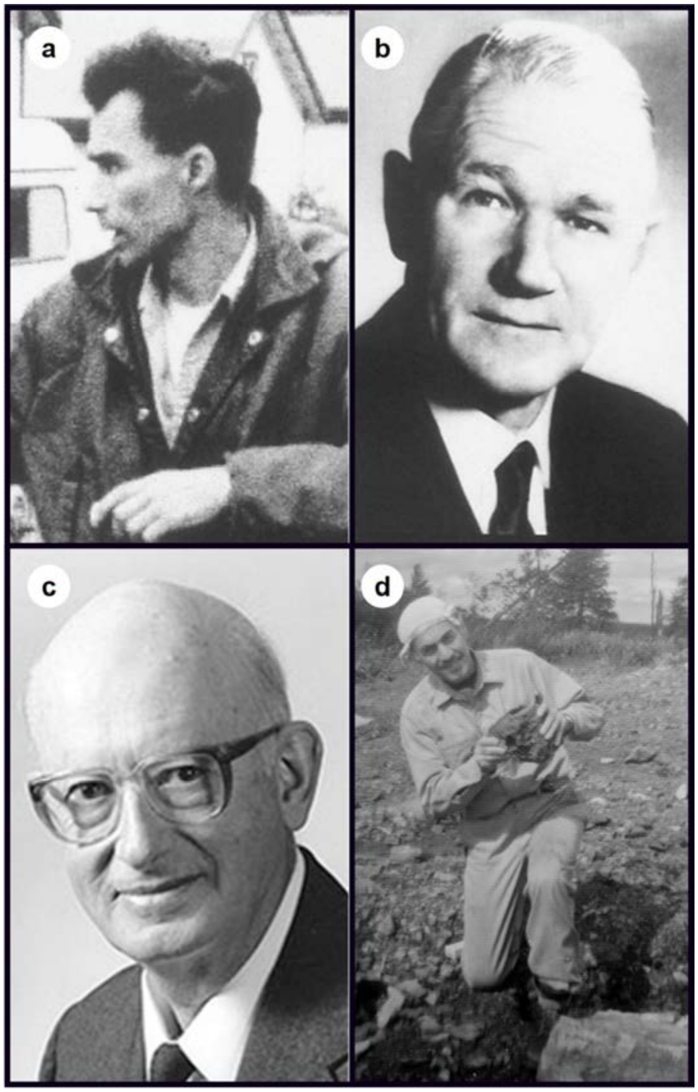

Figure 10. Photos: a) Dr. Cameron (Cam) G. Cheriton 1964, b) Dr. C. John Sullivan 1975, c) Dr. Richard L. Stanton c. 1985, d) Mr. Claude Willett 1975. The first two photographs are from the New Brunswick Department of Energy and Resource Development archives, the third is from the University of New England, and the fourth is courtesy of George Willett.

Two target areas were identified: one to the east called Number One and the other to the west called Number Two (Cheriton 1960). Work in the Number One area led to the discovery of the Armstrong A and B deposits in 1956, and the Rocky Turn deposit in 1957. The McMaster deposit was found between areas One and Two in 1957. At the time, the Number Two area was held by Bathurst prospector Fred J. Smith (unpublished Provincial files), so in 1955 Cheriton optioned Smith's Caribou property and changed the name to Anaconda-Caribou. Follow-up exploration identified drill targets and the first hole, completed in December 1955, intersected approximately l5 m of massive sulphide (Cheriton 1960).

In 1955, Kennco Explorations (Canada) Ltd. became active in the Bathurst area. Kennco examined geological data, aerial photographs and the government aeromagnetic maps, and three areas were selected for AEM surveys, to be conducted by Aeromagnetic Surveys Limited (Fleming 1961). Highly rated AEM anomalies were detected near Caribou, Clearwater and Murray Brook. Since, the anomaly at Caribou had already been acquired by Anaconda, only the Clearwater and Murray Brook anomalies were staked.

Ground follow-up in the Murray Brook area began with horizontal-loop and vertical-loop EM surveys to ground-truth the AEM anomalies. It was fortuitous that someone in the crew discovered copper-bearing float about 240 m south of the survey area at the end of the 1955 field season, because this justified further work the following year (Fleming 1961). The 1956 program used the newly developed stream-sediment geochemical methods of Hawkes and Bloom (1956). and led to the discovery of the Murray Brook gossan, which was drilled but none of the six packsack holes penetrated massive sulphide. Therefore, a ground EM survey was conducted over the gossan to help locate drilling targets. Drilling started in 1956 and the massive sulphide deposit was discovered about November 1st of that year (Fleming 1961).

Ground follow-up in the Clearwater Stream area began in the summer of 1955 with geological mapping and a horizontal-loop EM survey that located the AEM anomaly about 300 m south of its plotted position (Petruk 1957). In September, the ground anomaly was tested by two packsack drill holes, both of which intersected massive sulphide, discovering the Clear (later renamed Clearwater and then Chester) deposit.

In 1955, the President of Kennco Explorations (Canada) Limited, Dr. C. John Sullivan (Fig. 10b), who was an expatriate Australian, happened to read a paper by Dr. Richard L. Stanton (Fig. 10c), in which the author postulated that the Lower Paleozoic sulphide ores near Bathurst, New South Wales, had been deposited in a syn-sedimentary, volcanic island arc setting (Stanton 1955). Sullivan was so impressed with the apparent similarities to deposits in northern New Brunswick, that he wrote to Stanton and invited him to come to Bathurst (Stanton 1984). As a result of this letter, Stanton decided to accept a post-doctoral fellowship at Queen's University, which eventually enabled him to visit Bathurst for a month in the fall of 1956. During that time, he logged and sampled cores from the Brunswick deposits, which formed the basis for his paradigm-shifting papers (Stanton 1959, 1960a, b) on massive sulphide deposits.

The Consolidated Mining and Smelting Company of Canada Limited (Cominco), acquired ground in the Forty-Four Mile Brook and Nine Mile Brook areas in 1955 (Douglas 1965). However, a wedge-shaped area near Forty-Four Mile Brook was left open so it was staked by Bathurst prospectors (including Claude Willett) and named the Wedge property. A gossan outcrop was found and the property was optioned to Cominco. A ground EM survey was conducted in 1957 and one of the first holes drilled to test the EM anomaly intersected 33 m of massive sulphide, grading more than 4% Cu, thus discovering the Wedge deposit (Douglas 1965).

In the summer of 1954, Stratmat Limited, a wholly owned subsidiary of Strategic Materials Corporation, staked a group of claims (group 61) immediately north of the Heath Steele Mines property (Mowat 1957). Between then and the fall of

1956, airborne and ground electromagnetic surveys and soil geochemical surveys were performed. Diamond drilling in late 1956 discovered the "Group 61 Zone" (Johnston 1959), later renamed the Stratmat Main Zone (Dahn 1986).

Selco Exploration Co. Ltd. staked a group of claims in the Portage Lakes area in 1954, based on reconnaissance stream-geochemical work (Hawkes and Webb 1962, p. 331). Drilling of soil geochemical and ground EM anomalies discovered two small occurrences called the C-4 and C-5 zones, respectively. The property was optioned to New Jersey Zinc Exploration Co. (Canada) Ltd. in 1957 and the company drilled eight holes on the C-5 Zone and several holes on the C-4 Zone. Drilling of coincident soil geochemical and self-potential anomalies, just over 1 km to the south of the C-5 zone, discovered the Charlotte prospect in the fall of 1958 (Hawkes and Webb 1962, p. 327), which was later renamed the Restigouche deposit (*Mineral Occurrence Database*).

Eight other deposits were found in the 1950s, including the Captain, Devil's Elbow, Halfmile Lake North, Headway, Nepisiguit A, B, and C, and the Pabineau deposits (Table 1; Fig. 2). Except for Headway, which was found by prospecting, these deposits were discovered by electromagnetic methods.

1959–1973: The Flat Zinc Price Years

Of the six discoveries that were made during this period, most are satellite bodies to known deposits. Two of them (Table 1; Fig. 2) were found by Heath Steele Mines Ltd. by drill-testing IP anomalies. In 1964, the Heath Steele N-5 zone was found in the northern part of the mining lease (Hamilton and Park 1993). In 1966, the West Grid Zone was discovered approximately 1.5 km to the west of the Heath Steele ACD zone. The Stratmat Boundary zone was discovered by Cominco in 1961 on the Stratmat group of claims, which had been purchased from Strategic Metals in 1959 (Hamilton and Park 1993). The Canoe Landing Lake deposit was intersected by two packsack diamond drill holes in 1960 and further delineated by Baie Holdings Ltd. in its 1961–1962 drilling program. The discovery of the Louvicourt deposit is attributed to prospecting. In 1964, Lawrence Gray discovered gossan in the Nine Mile Brook area, when Route 430 was being constructed. He and partners L. Gamble and C. Smyth staked claims and subsequently optioned them to Louvicourt Goldfields Corporation. Drill testing of an SP anomaly led to the discovery of the Louvicourt deposit. Drilling of IP anomalies by Cominco led to the discovery of the Stratmat West Stringer zone in 1972.

1974–1988: The First Zinc Price Shift

During the 1970s and 1980s, new low-cost analytical and geophysical techniques sparked new discoveries, including five deposits (Table 1 and Fig. 2). In October of 1975, mineralized boulders were found by Sabina industries, 6 km southwest of the Brunswick No. 6 mine, while field checking an 'INPUT' airborne EM anomaly from a survey flown in 1974 by Questor Surveys Ltd. These boulders returned values of 5.2% Pb, 7.75% Zn and 38 g/t Ag (*The Northern Light*, October 15, 1975). Drilling in the fall of 1975 discovered the Flat Landing Brook deposit, which was subsequently optioned to United States Steel Corporation (Essex). In 1977, Consolidated Morrison found the Taylor Brook occurrence by drill-testing an AEM anomaly. With further drilling, this showing was upgraded to a deposit (Lutes1997). The Captain North Extension (CNE) deposit was discovered in 1978 by Sabina Industries and Metallgesellschaft Canada Ltd. (Whaley 1992). A stream geochemical anomaly and an IP survey delineated the drill targets that led to the discovery. In 1981, the Heath Steele C-North zone was found, approximately 550 m north of the ACD zone, by drill-testing combined ground Mag and EM anomalies. In 1988, the Stratmat S-l deposit was discovered by stratigraphic drilling at the Stratmat Central deposit to test the host sedimentary horizon between 230 m and 520 m below surface (Hamilton and Park 1993). One notable showing was found in 1975 by Claude Willett (Fig. 10d) while prospecting in the Nine Mile Brook area. He found high grade massive sulphide boulders that returned 4.55% Cu, 14.2% Pb, 8.55% Zn, and 487 g/t Ag (*Moncton Times*, August 7, 1975). A mini staking rush resulted; over 400 claims were staked by several companies and a bidding war ensued for Willett's claims. Ultimately, he optioned them to Price Company Limited (Newmont Mining Corp.) for a six-figure cash payment, an unheard-of sum for any property at the time.

1989–2000: The Second Zinc Price Shift

From 1989 to 2000, four deposits were discovered (Table 1 and Fig. 2). In 1989, deep stratigraphic drilling at the Brunswick No. 12 deposit discovered the Brunswick Northend zone approximately 1500 m north of the northern extremity of the main No. 12 deposit and 1100 m below surface (Hussey 1992). In 1992, Rio Algom Exploration optioned the Key Anacon property from Key Anacon Mines Limited. Drilling beneath the old mine workings later that year, intersected ore grade massive sulphide in hole 92-10 and hole 92-17 at vertical depths of 750 m and 450 m, respectively. In 1993, the Key Anacon East deposit was discovered by stratigraphic drilling beneath Carboniferous cover rocks, 1.5 km to the east-northeast of the Key Anacon deposit. The discovery hole (93-42) intersected 19. 9 m of 3.58% Pb, 7.86% Zn, 0.33% Cu and 78 g/t Ag within an 83 m massive sulphide intersection (Lentz and Langton 1993). In 1996, Noranda discovered the Camel Back deposit as part of a follow-up to the EXTECH airborne geophysical survey. A coincident Mag/EM anomaly 6 km southeast of the Caribou deposit was trenched and drilled in the latter part of the year. Hole 96-6 cut 17.9 m of massive and semi-massive sulphide mineralization, with a 4.3 m section returning 3.94% Pb, 8.95% Zn, 0.08% Cu and 41.9 g/t Ag, followed by a copper zone of 12.3 m grading 2.05% Cu. In 1999, Noranda found the Mount Fronsac North deposit following the discovery of a 50 m by 20 m gossan zone in a scarified forest harvest block (Graves and Mann 2000). The deposit is a 14 Mt sulphide accumulation at the contact between the Nepisiguit Falls and Flat Landing Brook formations (Walker and Graves 2006). After 1999, no new deposits were found.

IMPORTANCE OF THE BATHURST MINING CAMP TO MINERAL EXPLORATION

The Bathurst Mining Camp (BMC) was, and still is, important to New Brunswick and Canada for innovations in exploration methods and development of geological ideas.

Exploration Innovations

The BMC had several firsts vis-à-vis innovations in exploration. It was where:

- the first airborne magnetic survey was flown in New Brunswick;
- an airborne electromagnetic (AEM) survey was used for the first time to discover a VMS deposit (Heath Steele);
- ground gravity surveys were first used (starting in the 1950s) to screen electromagnetic (EM) anomalies;
- an airborne gravity system (Falcon) was first tested over a VMS deposit (Heath Steele), and that airborne gravity (Bell Geospace) was flown over an entire camp;
- directional drilling was used in a VMS environment for the first time;
- a 3D seismic survey was first used to discover the Halfmile Lake Deep deposit.

Geological Ideas

New geological ideas were developed and/or applied in the BMC. For example, the BMC was where:

- the syn-volcanic model of massive sulphide deposition was first applied (Stanton 1959) in Canada;
- it was determined that most VMS deposits formed in the Middle Ordovician during early-stage rifting of a submarine volcanic arc (floored by continental crust), which evolved into a Sea of Japan style back-arc basin situated on the eastern margin of Iapetus;
- it was determined that the present-day geology reflects amalgamation of various Ordovician rock units in a subduction/obduction complex (van Staal 1994) that formed in the Late Ordovician to Early Silurian;
- it was determined that the Miramichi Group is part of the Gander Zone and the Bathurst Supergroup belongs to the Dunnage Zone of the Canadian Appalachians.

ACKNOWLEDGEMENTS

We would like to thank the two Geoscience Canada reviewers (Robin Adair and Sean McClenaghan) for their helpful comments on the manuscript; most of their suggested changes have been incorporated in this final version. We would also like to thank Cees van Staal, Alex Wills and Wayne Goodfellow for providing their original CorelDraw files, from which Figures 2, 3, and 6 were prepared, respectively.

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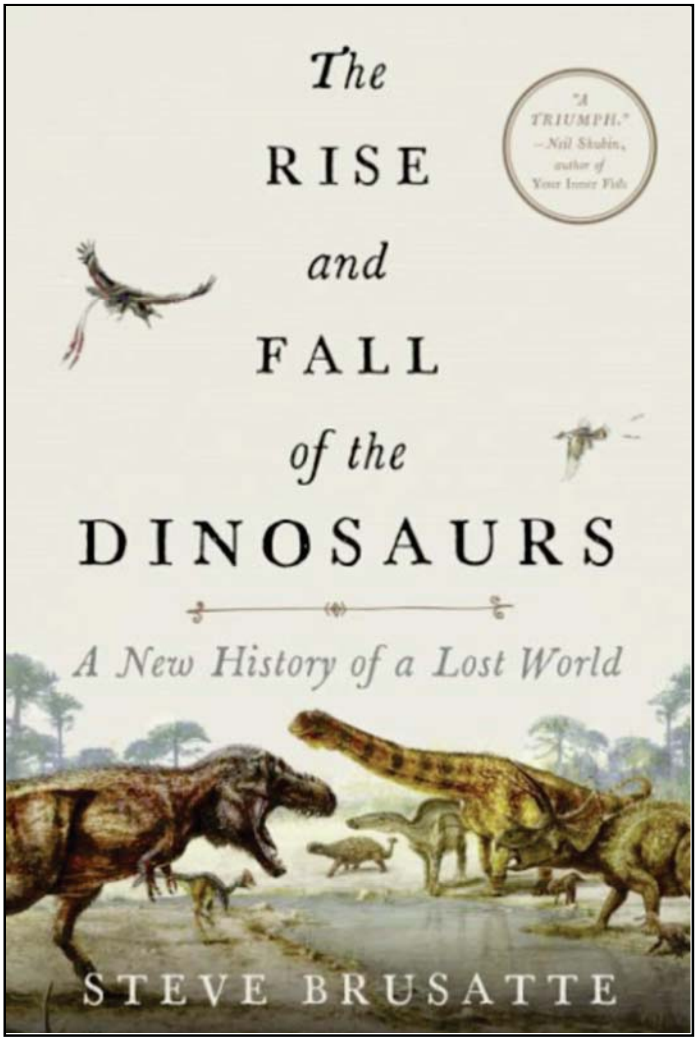
Received May 2019
Accepted as revised September 2019

REVIEW



The Rise and Fall of the Dinosaurs: A New History of a Lost World

Steve Brusatte

Published by: William Morrow/HarperCollins Publishers.
Published: 2018; 404 p.
Purchase price: $24.98 (CND; Paperback); Hardback, electronic and audiobook versions available.

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Serendipity lies behind this review. About three months ago, I spent a few days at an Airbnb owned by a couple with two teenage sons. Browsing their bookshelves, I was intrigued by the number of books on dinosaurs. There must have been at least a dozen, many well illustrated, ranging from non-scientific kids' books depicting herds of herbivores unsuspectingly munching lush foliage while being stalked by voracious bloodthirsty carnivores, to much more sophisticated tomes including encyclopedia complete with statistics and detailed reconstructions of hundreds of different dinosaur species. The books belonged to the teenagers, and it struck me that theirs was the generation that grew up with dinosaurs – dinosaur stickers, dinosaur images on their t-shirts and pyjamas, and dinosaur film stars in Steven Spielberg's 1993 blockbuster Jurassic Park (based on Michael Crichton's 1990 book of the same name). As in the movie, it appeared that the extinct animals had literally become alive for these two members of the post-millennial generation, many of whom apparently can name and recognize more than a dozen dinosaur species from a very young age. Moreover, the cornucopia of information in the encyclopedia was evidence that this surge of public interest in extinct animals was based on much new scientific research and information. Among the books was a hefty tome by Steve Brusatte entitled *The Rise and Fall of the Dinosaurs: A New History of a Lost World*, with resounding endorsements from The Washington Post ('a masterpiece'), The New York Times (bestseller), The Economist (recommended reading), The Smithsonian, The Times of London, Science News, Popular Mechanics, etc. (best popular science book of the year), and Scientific American ('the ultimate dinosaur biography'), among many others. After a short browse, I realized my education was in serious need of an update and I got my own copy a couple of weeks later.

Steve Brusatte is a young professor of paleontology at Edinburgh University, who in his own words has been a dinosaur nerd since a very young age. An American from small-town Illinois, he studied paleontology at the University of Chicago, Bristol University (UK), and Columbia University in New York, and this book combines his accumulated expert-

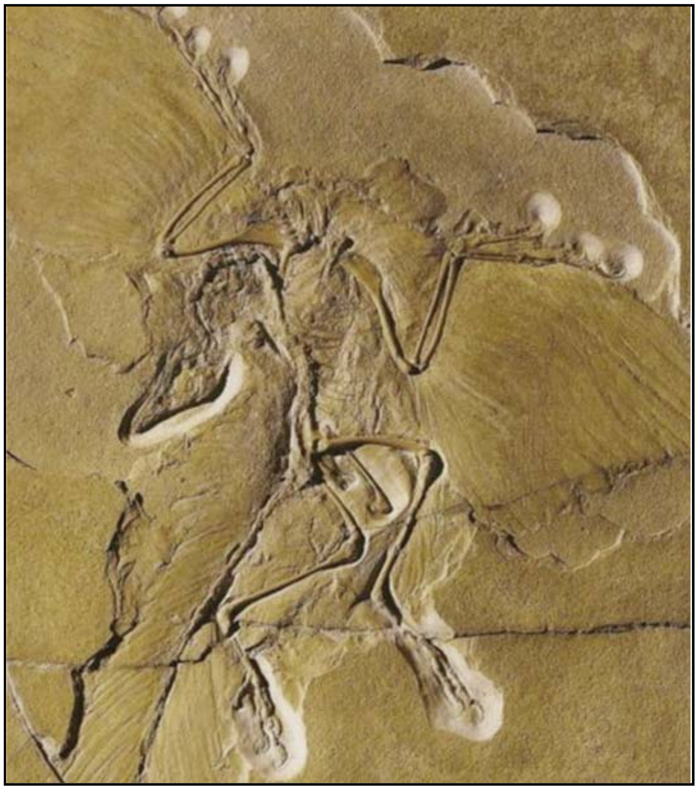
Geoscience Canada, v. 46, https://doi.org/10.12789/geocanj.2019.46.151 pages 155–157

Tyrannosaurus Rex, indisputably the best known of all dinosaurs, and fully deserving of its own chapter in Steve Brusatte's book. This impressive skeleton forms part of the exhibits at the Tyrrell Museum, in Drumheller, Alberta. Photo by A. Kerr.

ise and 'raw enthusiasm for dinosaurs' with his love of science writing and communication of popular science.

The storyline is bookended by two major extinctions, the end-Permian extinction at 252 Ma and the end-Cretaceous extinction at 66 Ma. The intervening 186 M.y. of the Mesozoic has yielded fossil evidence that the early, small bipedal protodinosaurs in the Triassic evolved from sprawling Permian precursors, dramatically diversified their anatomy, size, modes of locomotion and living, and flourished, becoming the dominant actors in the animal world on all the continents during the Jurassic and Cretaceous. The arrangement of the book is approximately chronological, from Triassic through Jurassic to Cretaceous, and much of the story is written through the first-hand experiences of the author as he travels the world to research critical issues in dinosaur evolution. Thus the reader gets a world tour, meets Brusatte and his mostly male buddies in the field and famous museums, and vicariously relives the excitement as they collect and document key dinosaur fossils, both bones and tracks, in Poland, Portugal, Romania, Italy, Montana, Wyoming, New Jersey, Scotland, China, Brazil, Argentina, and elsewhere. The field and museum experiences enliven the storyline and provide insight into the personalities involved — like Brusatte, many seem to be extrovert characters who caught the dinosaur bug at a young age, but Chinese expert Xu Xing, who has described and named more than fifty new dinosaur species, is an exception — he grew up poor in rural western China, won a scholarship to university in Beijing, and was told to study paleontology, a subject he had never heard of.

The book is about much more than evolving dinosaur anatomy and speciation from the small marginal Triassic *Eoraptor* and *Herrerosaurus* to the familiar Jurassic and Cretaceous giants such as *Brontosaurus*, *Tyrannosaurus* and *Triceratops*. The final assembly of Pangea took place in the Permian and the supercontinent had largely broken up by the end of the Cretaceous, and Brusatte is effective in incorporating paleoecology and showing how it and the changing arrangement of the continents affected dinosaur speciation and size, and the relative abundances of dinosaur lineages. There are also imaginative descriptions of the two extinctions that bookend the story, discussion of the climate on Pangea in the Triassic, climate change during the Mesozoic, and the dramatic sea-level rise in the Cretaceous — themes for our times. In addition, after a brief primer on the history of evolution and the origin of species, there is a compelling discussion of dinosaur 'missing links' in a chapter devoted to the origin of birds. The critical fossils, only discovered in the last twenty years or so in the Liaoning region of northeastern China, and more recently in Alberta, are exceptionally preserved specimens complete with impressions of feathers that are described by Brusatte as "*the fossils that help us untangle one of the biggest riddles of biology: how evolution produces radically new groups of organisms, with restyled bodies capable of new behaviors... what biologists call a major evolutionary transition*".



Although a great deal smaller than T. Rex, Archaeopteryx is every bit as important, as it provided the first clear evidence of a link between dinosaurs and birds. Reproduction at the Tyrrell Museum, Drumheller, Alberta. Photo by A. Kerr.

Also woven into the text are descriptions of the increasingly sophisticated methods employed by vertebrate paleontologists. For example, CAT scanning of dinosaur skulls is now routinely used to estimate brain volume, morphology and function; dinosaur body weight estimation, formerly little more than guesswork, now involves either algorithms based on dimensions of the femur, or more sophisticated species-specific techniques using photogrammetry to construct precise three-dimensional computer models of their skeletons to which muscles, internal organs, skin, etc. are added. These fleshed-out models can then be loaded into animation software and made to walk, run and jump — and finally, mechanically-plausible behaviour modes can be evaluated by stress analyses of bones and joints using finite element analysis. Another recent advance, statistical analysis of large datasets derived from measurements and observations of dinosaur fossils, has provided insight into dinosaur lineages. For example, it has been instrumental in showing that birds are theropods, a group of meat-eating dinosaurs that includes several famous names such as *Allosaurus*, *Tyrannosaurus*, and *Velociraptor*. Interestingly, in the context of the late 19th century subdivision of dinosaurs into saurischians (lizard-hipped dinosaurs) and ornithischians (bird-hipped dinosaurs), theropods are saurischians. The new research has shown that development of both feathers and the super-efficient bird lung, in which oxygen is extracted from the air during both inhalation and exhalation, occurred in the therapod lineage prior to the evolution of avian dinosaurs (yes; that terrible *Tyrannosaurus rex* was probably covered in a fuzz of proto-feathers).

This is a popular science book about the present 'golden age of discovery' in which about fifty new species of dinosaur are described in the paleontological literature each year and new research methods are constantly being added to the paleontologists' toolkit. The text is well-organized, fast-moving, and informative, each chapter covering a discrete topic while allowing the author scope for personal anecdote and historical vignettes. Chapter headings such as *The dawn of the dinosaurs, Dinosaurs become dominant, Dinosaurs and drifting continents, The king of the dinosaurs, Dinosaurs take flight*, and *Dinosaurs die out* indicate the accessible approach employed. For those wanting to follow up on details, there is an extensive list of sources at the end of the text, and the book also has an index. Frontispiece illustrations include a geological time chart, a dinosaur family tree, and plate tectonic reconstructions showing Pangea in the Triassic, and its breakup in the Jurassic and Cretaceous; and also each chapter is headlined by simple elegant black and white drawings of a dinosaur featured in the ensuing text. In addition, many chapters include photographs of the author and his colleagues in the field and laboratory, as well as images of exquisitely preserved dinosaur fossils. Brusatte obviously had fun writing this book, and frequently uses metaphors from popular culture to make his point; for instance this line: "*You could call T. rex the James Dean of dinosaurs: it lived fast and died young*". These analogies generally drew the desired reaction from me, but I found myself grimacing occasionally, for instance at the use of the word 'critters', anthropomorphic descriptions of 'plucky' proto-dinosaurs emerging from the Permian mass extinction, and the description of *Ankylosaurus* as 'stupid'. On the other hand, this metaphor for the Cretaceous–Paleocene boundary in Gubbio, Italy, where the abundant Cretaceous marine fossil record gives way above the iridium anomaly to a complete absence of fossils in the early Paleocene, hit the sweet spot for me: "*Walter* [Alvarez] *was observing a line between life and death. It's the geological equivalent of listening to those last few moments on a cockpit voice recorder before it gives way to silence*".

Did I mention serendipity? As a moderately sized mammal, I would not be writing this review if a large bolide had not impacted Earth at the end of the Cretaceous, forming the Chicxulub crater and that iridium anomaly and wiping out the non-avian dinosaurs and much else besides, thereby making way for the rise of the mammals in the Cenozoic.

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GEOSCIENCE CANADA AND THE GEOLOGICAL ASSOCIATION OF CANADA ARE GRATEFUL TO THE CANADIAN GEOLOGICAL FOUNDATION FOR THEIR FINANCIAL SUPPORT OF THIS JOURNAL.

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SERIES

REVIEW

2019 VOLUME 46 GSCNA7 46 101-158
PRINT EDITION ISSN 0315-0941
ONLINE EDITION ISSN 1911-4850