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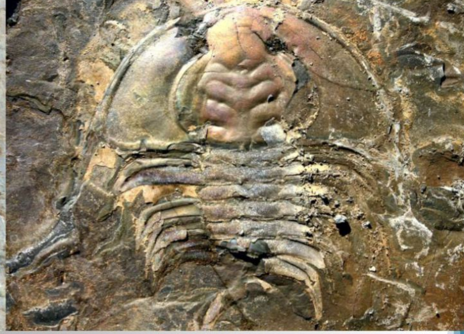

GEOLOGICAL GUIDE TO THE BIRD COVE REGION, GREAT NORTHERN PENINSULA

I. Knight and W.D. Boyce

Open File NFLD/3239

St. John's, Newfoundland
July, 2015

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SAFETY CAUTION

Many of the localities discussed in this guide can be visited along roadsides, in quarries and along the shore. In each case, care should be exercised to avoid injury by moving vehicles, falling rocks (use hard hats if possible in quarries) and rough seas. Suitable walking boots and rain clothes are also recommended. Best to be prepared.

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Cover: Meany's Point mounds, looking east across the mouth of Long Bottom toward Bird Cove. Fossils include Early Ordovician gastropods and Early Cambrian trilobite Olenellus transitans *(Walcott, 1910).*

Newfoundland Labrador

Natural Resources

Mines

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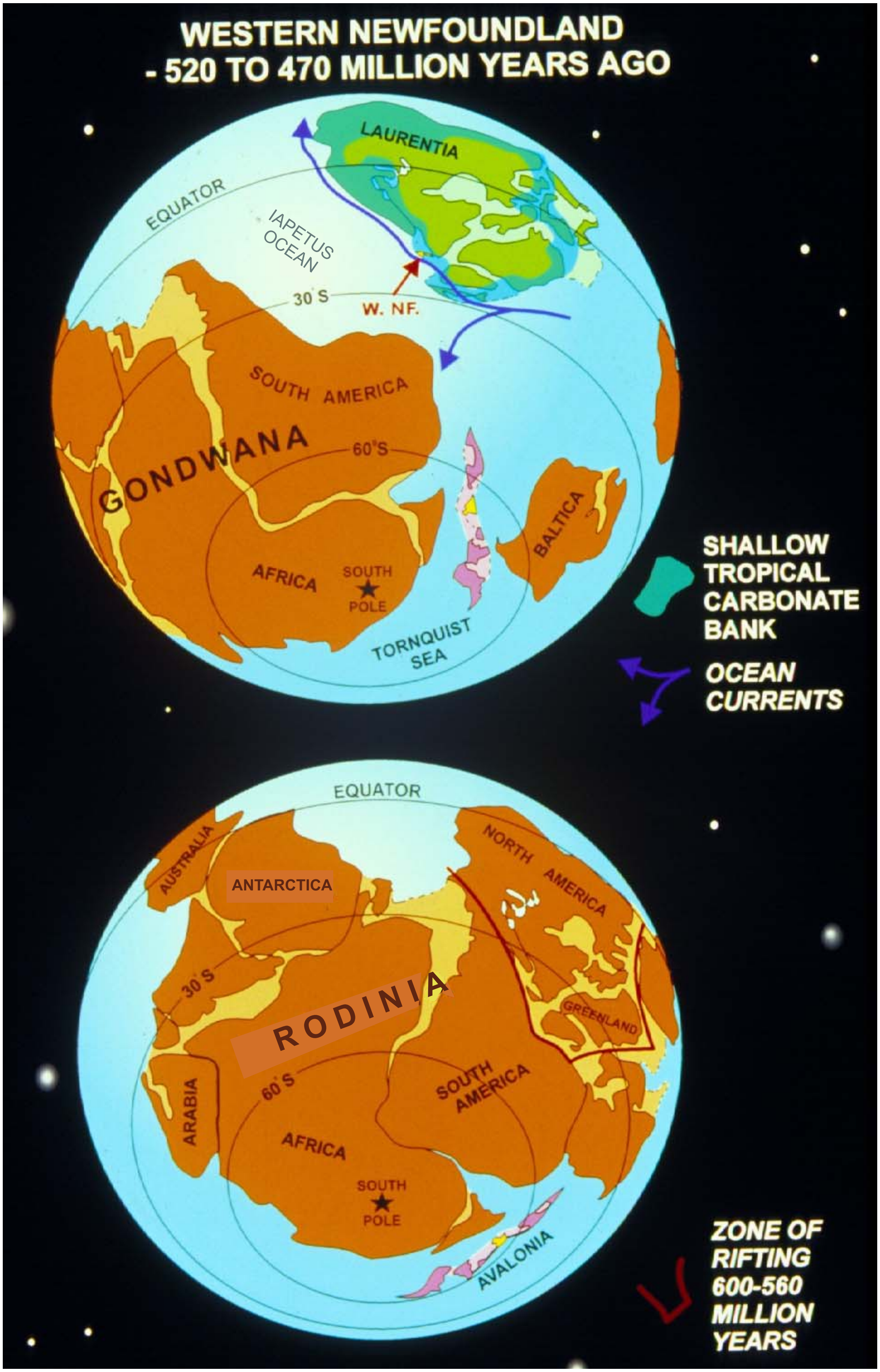

Earth, its continents and oceans, between about 600 and 500 million years ago. The lower globe shows a supercontinent **Rodinia** that split into various parts about 600 million years ago. **Laurentia** (upper globe) rifted apart and drifted gradually north, away from the supercontinent, creating the Iapetus Ocean. Western Newfoundland (W.NF.) is indicated.

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ABSTRACT

Geologically contrasting physiographic terrains dominated by Cambrian–Ordovician siliciclastic and carbonate rocks underlie the Bird Cove region on the Great Northern Peninsula. Gently dipping, Lower Cambrian sedimentary rocks resting unconformably on truncated Grenvillian crystalline basement rocks underlie the fault-bounded uplands of Mount St. Margaret, the Highlands of St. John and the northern end of the Long Range Mountains. The late Early Cambrian Labrador Group comprises basal non-marine to marine sandstones overlain by deep- to shallow-water shelf limestone, shale and sandstone. These rocks reflect marine flooding and later shelf progradation along Newfoundland's strand of the Laurentian margin. Often a richly fossiliferous succession, it is host to shelly fossils, numerous trace fossils and archaeocyathid patch and barrier reef complexes.

West of the highland terrain, low-lying coastal lowlands of St. John, St. Margaret and Ste. Genevieve bays, including the Dog and New Ferolle peninsulas are underlain by essentially flat-lying, Upper Cambrian and Lower Ordovician rocks. The Port au Port and St. George groups preserve parts of a long-lived (40 Ma), platformal carbonate succession deposited in warm tropical shallow seas. The modern day Bahama Banks is the best modern analogue with which to compare the ancient succession. Microbial mounds associated with mudcracked dolostone displaying salt-crystal pseudomorphs suggest a platform subject to high salinities in the Late Cambrian. By contrast, algal and burrowed limestone plus gastropods and other fossils support near normal salinities in the Early Ordovician.

Reverse faults, the Castor Pond, Ten Mile Lake and the Plum Point faults, mark the western limit of Appalachian deformation in western Newfoundland. Steeply dipping strata against the faults contrast with the gently dipping and folded rocks in between.

Blanketed by thick ice sheets at the height of the Wisconsin Glaciation (~18 000 yrs BP), the postglacial history of the area after 13 000 yrs BP is one of ice retreat, marine drowning to 140 m above present sea level at about 12 000 yrs BP and then gradual emergence of the area in response to isostatic rebound. Numerous raised beaches overprinted by karst around the shorelines of the Dog Peninsula reflect the later stages of this rebound.

Chert, mostly in Ordovician rocks, may have provided ancient Maritime cultures, which inhabited the area, a local source of stone to manufacture their implements. The postglacial raised beaches, low elevated ridges and natural harbours of the Bird Cove area provided sites for settlement.

INTRODUCTION

Bird Cove and the Dog Peninsula lie at a pivotal point along the Great Northern Peninsula (GNP) of western Newfoundland, as the long mountainous spine of the peninsula, inland to the east, abruptly ends. The coastal lowlands of woods and marshes, which fringe the western edge of the Long Range Mountains for over 300 km north of Rocky Harbour, widen to a broad, 60-km-wide lowland rarely rising more than ~60 m (200’) north of the mountains. A significant part of this northern lowland is a classic carbonate barren that stretches east across the Northern Peninsula from the Dog Peninsula to Hare Bay and Canada Bay, and continues north to Cape Norman, over 70 km away (Figure 1).

This jarring change of physical geography echoes the underlying rocks and geological history of the region that stretches back over 1.5 billion years (*see* Figures 1 and 2). The mountainous spine, with its rugged glacial valleys incised down below the rocky mountain plateau, is mostly carved in ancient crystalline rocks, principally granite and gneiss, 900 to 1600 million years old. In contrast, the low-lying barren is a terrane of flat-lying to gently dipping, much softer and more readily eroded, stratified carbonate rocks (limestone and dolostone). The carbonates were once (half a billion years ago) part of a vast tropical shallow-marine continental shelf, called the **Great American Carbonate Bank**, that fringed the ancient forebear of North America, **Laurentia**. The precipitous mountain fronts that mark the transition from the upland plateau to the coastal lowlands coincide, almost without exception, with inactive fault lines.

The Highlands of St. John (HSJ, Figure 3; maximum height 620 m) and Mount St. Margaret (MSM; 275 m) essentially interrupt the coastal lowlands between Hawke Bay and the Dog Peninsula. These long, flat irons, or cuestas form a distinct frontal range just northwest of the Long Range Mountains’ spine as it reaches its northern terminus. The highlands, which reach within a few kilometres of the sweep of St. John, St. Margaret and Ste. Genevieve bays, are formed, in part, of strongly erosion-resistant, essentially flat-lying, siliciclastic sedimentary rocks resting upon older crystalline rocks, all of which are faulted against the limestone and dolostone of the coastal lowlands.

This Open File report about the Bird Cove area is written with the object of providing an educational resource for both professional and private citizens interested in its geology. It updates and replaces a similar earlier report (Knight and Boyce, 2003).

GLACIAL HISTORY

Low terraces and ridges, a few metres above present sea level, made from locally derived well-rounded and angular dolostone and incompletely carpeted by low-growing shrubs and flowers snake along the coastline of the Dog Peninsula. These features, mostly ancient beach ridges stranded above the present strandline, date from about 5000 to 2000 yrs BP. Older beach deposits are perched along the southern slopes of the Highlands of St. John many 10s of metres above present sea level. All these features indicate that over the last 10 000 years or more, the physical geography and landscape of the area has changed remarkably as isostatic rebound affected this part of the Great Northern Peninsula, following retreat of the great ice sheets of the last Wisconsin ice age.

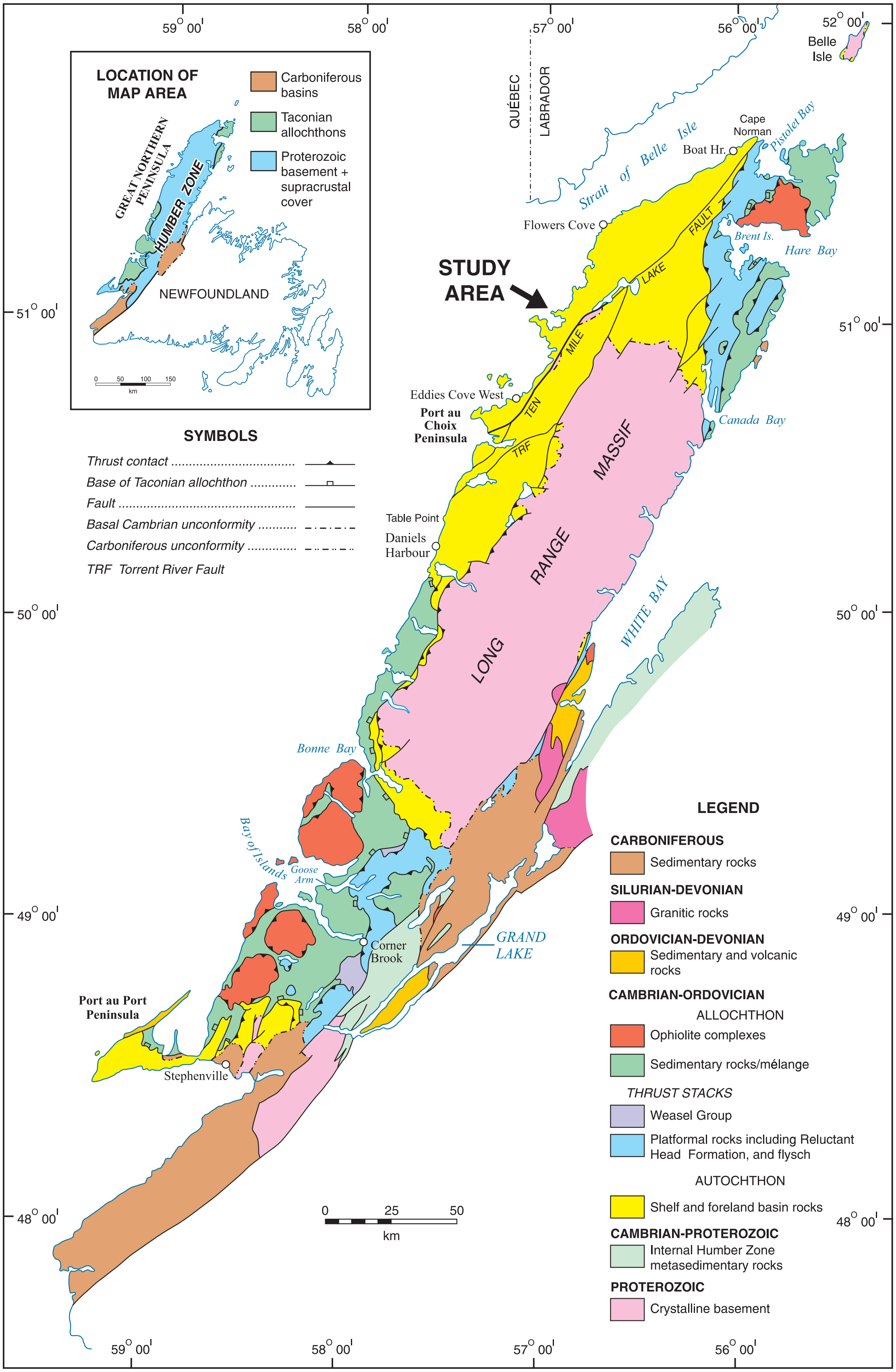

Figure 1. *Geology of western Newfoundland.*

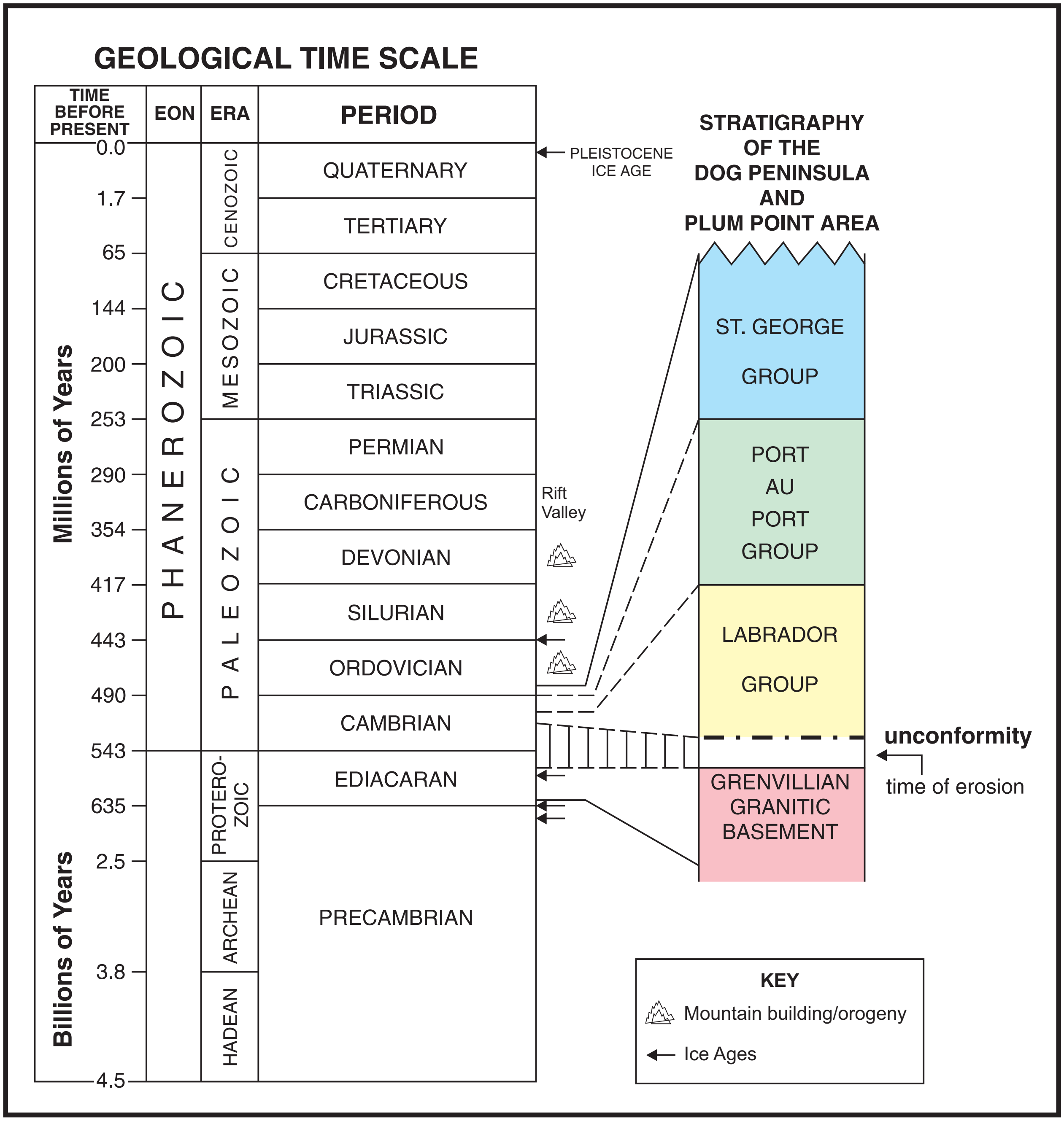

Figure 2. *Geological time scale, and a simplified stratigraphy of rocks of the Bird Cove area.*

The process of isostatic rebound may seem counterintuitive when looking at or falling on the hard rocks of the area. But although, the Earth's crust appears strong, rigid and uncompressible, on the grand scale of the Earth, crustal rocks have flexibility as well as rigidity and strength. Also, the Earth's crust is essentially floating on semi-molten, hot rocks deep in the Earth that allow the crust to regionally depress when a load is applied on top. Thick ice sheets, sometimes many kilometres thick, provide such a weight to bear down on the underlying crustal rocks.

Figure 3. *The North Summit of the Highlands of St. John as seen from the Viking Highway (Route 430). Well-bedded, almost flat-lying sedimentary rocks of the Cambrian upper Labrador Group are exposed in the mountain's scarp face. The narrow saddle between the scarp and the small hill (arrow) marks the trace of the Ten Mile Lake Fault.*

A simple analogy for isostatic rebound is provided by a spring board used in diving competitions, and another by applying a weight to a trampoline. The diving spring board fixed at one end and without a diver on it protrudes straight out. When a diver (the ice sheet) walks out to the end of the board, the board (Earth's crust) bends downward away from the fixed end and remains so for as long as the weight of the person is applied. When the diver walks off the board (ice sheet melts), their weight is removed and the board returns to its original horizontal attitude. The board is both rigid and flexible and responds temporarily to the weight by bending, *i.e.*, loading. When the load is removed the board returns to its original shape, *i.e.*, isostatic rebound.

A weight in the middle of a trampoline provides a slightly different insight into the effect of an ice sheet sitting on the Earth's crust. A gymnast standing in the middle of a trampoline creates a basin-like depression that is much larger in area than the footprint of the gymnast. Likewise, the vast weight of a thick ice sheet weighs down the Earth's crust to form a depression that can extend far beyond the edge of the ice sheet. The thick ice sheet covering Labrador and northern Newfoundland effectively depresses the Earth's surface to the north of the Bird Cove area, while at the same time differentially depressing a much wider zone of the Earth's crust, affecting shorelines far from the ice centre and beyond the edge of the ice sheet. Unlike the instantaneous change of shape of a trampoline when a weight is applied and withdrawn, the isostatic, geological process of ice loading, in response to the growth and retreat of ice sheets, occurs at a snail's pace and will affect a huge region of a continent over tens of thousands of years.

The following is a brief summary of the complex glacial history that shaped the modern landscape of the Bird Cove area. Before deglaciation and at the last glacial maximum (~18 000 yrs BP), the area was blanketed by a vast ice sheet, the confluence of the northern **Laurentide Ice Sheet** (LIS) of Labrador and the **Newfoundland Ice Cap** (NIC, Figure 4A; Grant, 1992, 1994; Shaw *et al.*, 2006; Putt *et al.*, 2010). The ice streamed southwest into the Gulf of St. Lawrence, merging with other regional ice streams to pass through the Cabot Strait and out to the edge of the present day continental shelf (Figure 4A). However, the ice sheet collapsed by 13 000 yrs BP, calving icebergs into a seaway named the **Goldthwait Sea** (essentially present day Gulf of St. Lawrence) as it rapidly retreated (Grant, 1992; Putt *et al.*, 2010). Ice escaping westward from the ice cap over the Northern Peninsula into the Gulf of St. Lawrence carved northwest-trending valleys such as Castor River and Doctors Brook throughout the glacial maximum and as the ice cover shrank. The Highlands of St. John, once ice covered at the height of glaciation, emerged as large rock masses (**nunataks**) through the Northern Peninsula ice cap (Grant, 1992). Numerous linear northwest-trending ridges of glacial debris, called **de Geer moraines**, are found throughout the northern coastal lowlands and western part of the carbonate barren (Figure 4B). These moraines show that the ice margin front, as it retreated in a northeast direction with the melting of the Northern Peninsula ice cap, oscillated back and forth seasonally or longer in a frigid two-step to deposit linear piles of glacial debris. Each de Geer moraine marks these temporary advances of the ice front during its overall retreat (Grant, 1992; Shaw *et al.*, 2006; Putt *et al.*, 2010).

Sea level rose across the region because of both global ice sheet melting and the depression of the region by the weight of the LIS. The two regional ice bodies (LIS and NIC, Figure 4A), separated by about 12 000 yrs BP and the margin of the floating ice sheet that lay close to the present shore of the Bird Cove area at 13 000 yrs BP (Figure 4A), withdrew to the edge of the Long Range Mountains and the coast of Labrador (Figure 4B; Grant, 1992, 1994; Putt *et al.*, 2010). This effectively drowned the northern half of the Great Northern Peninsula by as much as 140 m (just below 460 ft) over a few thousand years and the northern lowlands and carbonate barrens formed the seabed of a 40- to 60-km-wide seaway, part of the Goldthwait Sea (Figure 4B; Grant, 1992; Putt *et al.*, 2010). A brief cooling period at about 11 500 yrs BP, however, saw the Northern Peninsula ice cap advance north into the seaway to form a long arcuate moraine that is preserved at Ten Mile Lake and extends 30 km to the east (Figure 4B).

Isostatic rebound of the area previously regionally depressed by the vast weight of the 2.5 km thick LIS (M. Batterson, personal communication, 2011), however, forced steady but decelerating emergence of the area from about on average 2.3 m per century pre-10 000 yrs BP to 13 cm per century since 5000 yrs BP (Bell *et al.*, 2005a, b). Evidence of this submergence is shown by stranded beach deposits of large boulder and pebbles at approximately 80 to 100 m (~240 to 328’) above present sea level on the gently terraced hillsides of the southern Highlands of St. John (Knight, 1991; Bell *et al.*, 2005a). Encrusting marine organisms on cliff rock faces at lower levels in the Hawke Bay area, still far inland of the present coast, provide support for the widespread submergence of the area (Bell *et al.*, 2005a).

At about 12 000 yrs BP, the time of deepest drowning of the area, the shoreline likely lay at the base of the scarp face of the Highlands of St. John and Mount St. Margaret (Figure 4B). Using the 150 m contour as a guide it is likely that the sea reached as far inland as the Long Range Mountain spine along the valley of Leg Pond near Castor River and that the shoreline of the

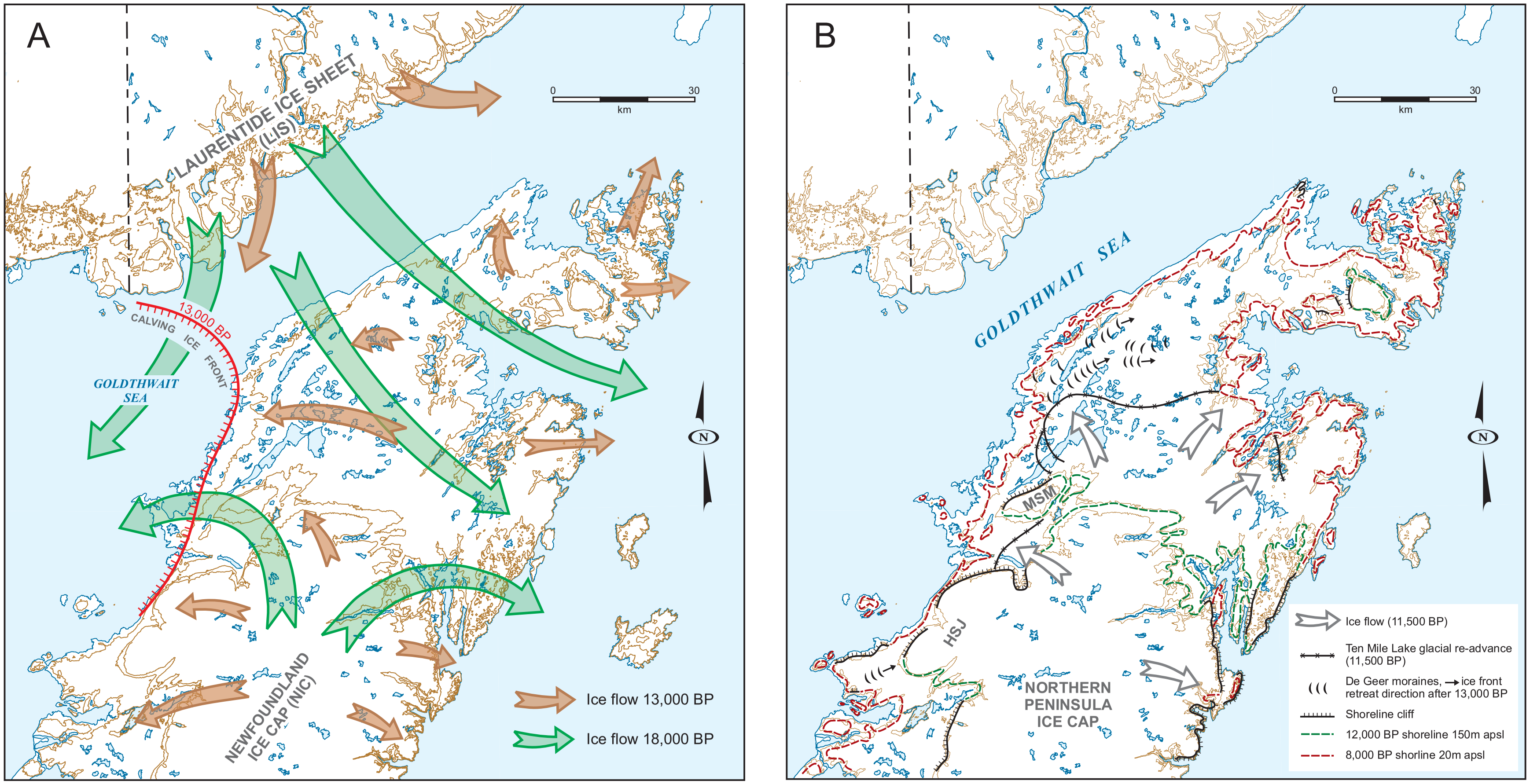

Figure 4. *Simple maps of the glacial history of the northern part of the Great Northern Peninsula during the last 20 000 years, based on Grant (1992), Shaw* et al. *(2006) and Putt* et al. *(2010). A) Map of ice flows during late glacial maximum (~18K yrs BP) and later collapse of the ice sheets (13K yrs BP); B) Approximate shorelines at 12K and 8K yrs BP as isostatic rebound dominated the area. The limit of the Ten Mile Lake Glacial Re-advance at 11.5K yrs BP is also shown. HSJ - Highlands of St. John; MSM - Mount St. Margaret.*

Goldthwait Sea snaked eastward along the northern edge of the Long Range Mountains to meet with the North Atlantic near Canada Bay. Freed of the burden of the NIC and the rapid retreat of the much larger continental LIS northward away from the Strait of Belle Isle, emergence of the Great Northern Peninsula gathered pace so that by 8000 yrs BP the shoreline was at approximately 20 m (about 65 ft) above present sea level.

The record of emergence of the Bird Cove area after 5000 yrs BP, is partly preserved by a series of raised beaches left stranded as the sea level fall continued, and it is likely that the coastal lowlands of the area, with its attendant bays and harbours, are little changed for the last 3000 years. Many of these stranded beach deposits on the Dog Peninsula (Figure 5A and B) are found at the western end of the peninsula. Some raised beaches lie facing the open water of the Gulf of St. Lawrence and comprise small, fairly well-sorted angular pebbles and rubble (Figure 5A). Raised beaches preserved in relatively storm-sheltered areas consist of angular dolostone rubble (Figure 5B and C) which shows an interesting contrast between the pock-marked and spiky, buff- and yellow-weathering, upper surface and the smoother, although somewhat rotted, brown undersurface of the beach rubble. The upper surface is typical of karst weathering on surfaces long exposed to dissolution below shallow, acidic and humus-rich soils and weathering by rain, sea spray, and melting snow. The angularity and contrast in weathering of the surfaces of the rubble indicate that the beaches were rapidly abandoned as sea level fell.

Since ancient cultures were strongly reliant on the sea and its resources, it is likely that archeological sites on the Dog Peninsula reflect the emergence (approximately 3 m) of the shoreline over this last 5000 years (*see* Bell *et al.*, 2005b).

HISTORY OF GEOLOGICAL STUDIES

The first geological research in northwest Newfoundland began in 1861 when the geologist James Richardson surveyed the coastline of the Great Northern Peninsula for the Geological Survey of Canada. He examined and collected rocks and fossils during a broad regional survey of the west coast of the peninsula. His fossil collections and geological observations were published in the *Geology of Canada* (Logan *et al.*, 1863), and in a treatise on fossils by Billings (1861, 1865). In the late 19th century and early 20th century, Alexander Murray and James P. Howley of the Geological Survey of Newfoundland undertook similar surveys and mistakenly showed on their maps that Silurian rocks underlay much of the coast and coastal plain of the Great Northern Peninsula, near Bird Cove (Howley, 1907).

During the early part of the 20th century, expeditions, funded principally by the Peabody Museum and Yale University in the United States, added to the geological understanding of the coast (Schuchert and Dunbar, 1934). The geologists would have doggedly walked and sailed the coast linking the rocks and fossils to those known elsewhere in Canada and the USA. Geological mapping of the area of St. John Bay and the adjoining highlands began with the expansion of the Geological Survey of Newfoundland under the leadership of David Baird in the 1950s. Staff and students from Yale University, namely Sam Nelson and Hank Woodard, were sent to map the area (Nelson, 1955; Woodard, 1957). The latter scaled the Highlands of St. John striking a base camp at the foot of the Long Range Mountains near Doctors Brook with the support of the inhabitants of Barr'd Harbour.

Figure 5. *Raised beaches at the west end of the Dog Peninsula. A) Raised beach of yellow-weathering, fairly well-sorted, small but angular dolostone pebbles derived from erosion of the underlying Cambrian dolostone. The beach faces toward the north-west and the Gulf of St. Lawrence; B and C) Angular blocky rubble of a raised beach on the southern lee side of the peninsula. Note the extreme angularity of the debris suggesting no erosion after breakup of the local dolostone probably by frost action. C is a closeup of the dolostone blocks showing characteristic pitting due to dissolution of the soluble dolostone. Note the weathering pattern on the large block suggests the original rock was part of a stromatolite mound.*

Although the area is covered by a regional geological map of the area by the Geological Survey of Canada (Bostock *et al.*, 1983), the most recent geological maps and studies of the Cambrian and Ordovician rocks of the region are principally the work of the Geological Survey of Newfoundland and Labrador, work that began in the late 1970s and continues to this day.

BASIC ELEMENTS OF PLATE TECTONICS

The present understanding of the geology of the Bird Cove area is a reflection of the late 20th century understanding of the dynamic geological mechanism of plate tectonics (*see* Geological Road Map of Newfoundland and Labrador, 2009 and Eyles and Miall, 2007 for good illustrated reviews). The theory of plate tectonics states that the Earth's crust is a series of large and small, mobile but rigid continental and oceanic plates that float upon the Earth's upper mantle. The plates are never static but move imperceptibly across the globe at a few centimetres each year. The interiors of continental plates are generally stable cratons formed of ancient rock formations that hide a complex geological history including many of the elements discussed below. Stable cratons generally are modified by long periods of weathering and erosion and provide sediment to basins mostly around their margins. The Canadian Shield is one such craton that formed by the accretion (tectonic suturing/stitching together) of several ancient Precambrian continental masses all with their own complex geological histories (Eyles and Miall, 2007).

Plate margins are critical geological boundaries. There, plates interact with each other. They are, therefore, the loci of much of the tectonic activity of our planet, be it earthquakes, volcanic eruptions, mountain-building, or sediment deposition from the erosion of the cratons and fringing mountain belts. For example, as new volcanic crust is created at mid-oceanic ridges, *e.g.*, Iceland, the expanding oceanic plates push older volcanic oceanic crust away from the ridge; continents attached to the spreading oceanic plate are also carried along, *e.g.*, the Atlantic Ocean and the North and South American continents. If this process of extension and generation of new crust begins in the middle of continents, long complex rift valley systems *e.g.*, the East African rift valley form. Sedimentary basins in such continental rifts are characterized by thick successions of river and lake sediments as well as local volcanic activity and strong earth tremors. Late Proterozoic non-marine and locally marine sedimentary and volcanic rocks in the Canada Bay, Belle Isle and southeastern Labrador were deposited in such rift basins (*see* below).

As continental rifting continues, continents split apart and new seaways and often deep but narrow ocean basins form *e.g.*, the Red Sea. Consequently, narrow rift valley basins can give way to true continental margins along which evolve broad, but shallow, continental shelves. The shelves build out over long spans of geological time into the newly formed deep ocean basins that widen as extension continues. A continental shelf located in tropical seas far from or cut off from rivers tends to be dominated by carbonate rocks (limestone and dolostone), for example the modern Bahama Banks (Figure 6) and the Great (Australian) Barrier Reef. Shelves located in cooler climatic zones and fed by rivers tend to be built up mostly by prodigious quantities of terrigenous sand and mud, *e.g.*, Grand Banks of Newfoundland, although limestones are still known to form in some modern cool water settings. The sedimentary rocks of the Bird Cove area formed along a tropical continental margin 500 million years ago (Figure 2).

Figure 6. *An aerial view of the tropical carbonate shelf of the Bahama Banks that provides a modern analogue for the carbonate rocks of western Newfoundland in the Cambrian and Ordovician periods. Tidal flats occur behind and left of the island, the large white area is an oölitic sand shoal associated with a tidal delta. A shallow lagoon occurs in the background and a shallow shelf in the foreground.*

Plates also converge and collide with each other. In some cases, plates are forced to move sideways against each other along major fault systems *e.g.*, the San Andreas Fault system of California and the Alpine Fault of the South Island of New Zealand. Major earthquakes affect these transform margins and sedimentary basins develop above the plate boundaries. Basins along an oceanic margin fill with marine sediments and are often prolific sources of hydrocarbons as in southern California. However, if the plates are continental, the basins are likely to be filled with non-marine continental sediment deposited by rivers (fluvial) and in lakes (lacustrine) across the floors of land locked basins. Two ancient land locked basins of this type occur in western Newfoundland. The Deer Lake and Bay St. George basins (Hyde, 1995; Knight, 1983) date to the Carboniferous (~360 to 310 million years ago, IUGS, 2009); modern examples include the Dead Sea rift valley in the Middle East.

Convergent margins where plates essentially oppose one another subject rocks to intense compression as one plate is forced to dive under the other, a process called subduction. The descending plate carries volcanic and sedimentary rocks and water deep into the Earth where they are transformed by deformation, metamorphism and by melting to provide a variety of copious magmas that feed volcanoes in the overriding plate as well as huge intrusive bodies of plutonic rock, such as granite. Sedimentary rocks are scraped from the seafloor and plastered (accreted) by various processes to the overriding plate. In some cases, enormous blocks of seafloor can be thrust onto continental margins to form ophiolites such as the Lewis Hills, Blow-Me-Down and Tablelands of Gros Morne National Park, a process called obduction. When oceanic plates converge in mid ocean, ocean trenches and volcanic-island arcs are commonly formed, as for example the Pacific rim of fire. Where oceanic plates subduct beneath continents *e.g.*, Pacific plates beneath the west coasts of North and South America, the convergent zone is characterized by long, linear to sinuous mountain belts epitomized by a mix of deformed oceanic and continental sedimentary, volcanic and metamorphic rocks, volcanic provinces and complex episodes of igneous intrusions of all varieties including granites.

It is also possible for continents to collide as for example the India subcontinent with central Asia and the northern edge of Africa with Europe. Spectacular alpine mountain ranges are characteristic of these wrestling continents that are essentially sutured together, perhaps the beginning of a super continent. Rocks are often thrust faulted over great distances, deformed into huge folds and metamorphosed deep in the zone of collision. These mountain belts generally have extreme topography, deep roots, earthquakes and persistent, often catastrophic erosion that produces copious amounts of sediment to be carried by river systems, into nearby, and often distant oceans and basins.

The mountains of these convergent zones, of which the Appalachian Mountains are a prime example, preserve the history of complex **orogeny**. Distinct orogenic phases, distinguished from one another by crosscutting and stratigraphic rock relationships and radiometric age dating of volcanic, intrusive and metamorphic rocks, are named to allow informed discussion of the mountain building process. In the Appalachians of Canada, four orogenies built the Appalachian Mountains, the Ordovician **Taconic Orogeny**, the Silurian **Salinic Orogeny**, the Devonian **Acadian Orogeny** and the Carboniferous **Alleghenian Orogeny**.

The tectonic amalgamation of many geological plates forms super- or mega-continents. This is known to have occurred repeatedly through the approximately 3.5 billion years of the Earth's geological history only for them to be later broken apart by plate tectonic processes. Deposits of rifted continental margins and continental shelves are therefore easily incorporated in mountain belts created by convergence, essentially the geological story of the Bird Cove area.

GEOLOGY OF THE BIRD COVE AREA

GRENVILLIAN BASEMENT ROCKS: PROTEROZOIC GLOBAL WANDERING LANDLOCKED IN THE SUPERCONTINENT RODINIA

The Bird Cove area lies at the frontal edge of the Newfoundland Appalachian Mountains essentially in a zone where the mountain belt abuts with the cratonic Canadian Shield. Although the rocks of the Bird Cove area consist predominantly of sedimentary rocks of the early Paleozoic, spanning a time interval between ~520 and 480 Ma, the oldest rocks in the area are **Proterozoic** (900 to 1600 Ma). The Proterozoic rocks outcrop in the Long Range Mountains and occur as a narrow zone mapped at the foot of Mount St. Margaret to as far south as the North Summit of the Highlands of St. John (Figures 7 and 8). Rocks in this narrow zone are exposed along Route 432

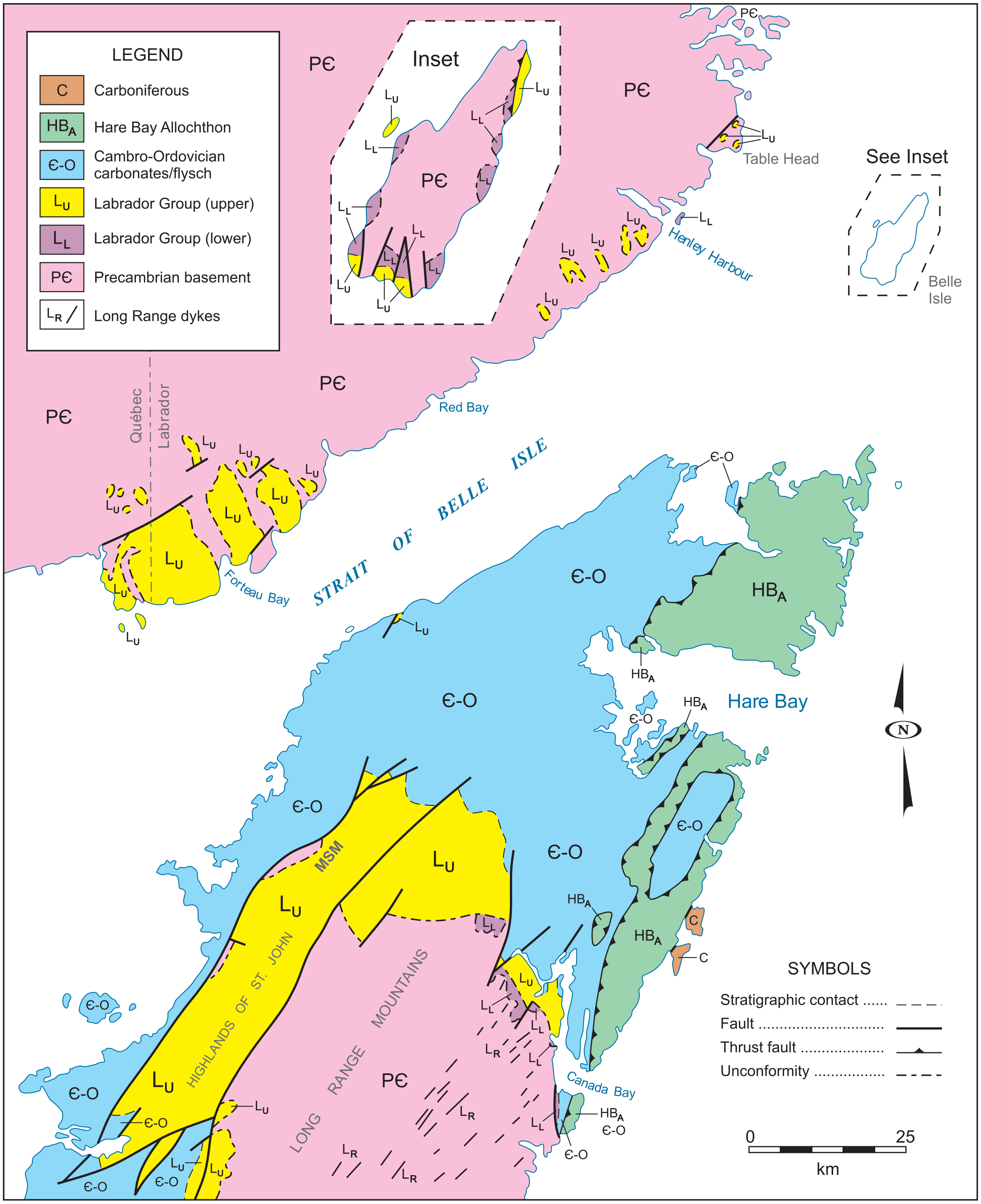

Figure 7. *Geology map showing the distribution of lower and upper Labrador Group and Long Range dykes on the Great Northern Peninsula (based on Gower* et al.*, 2001). MSM - Mount St. Margaret.*

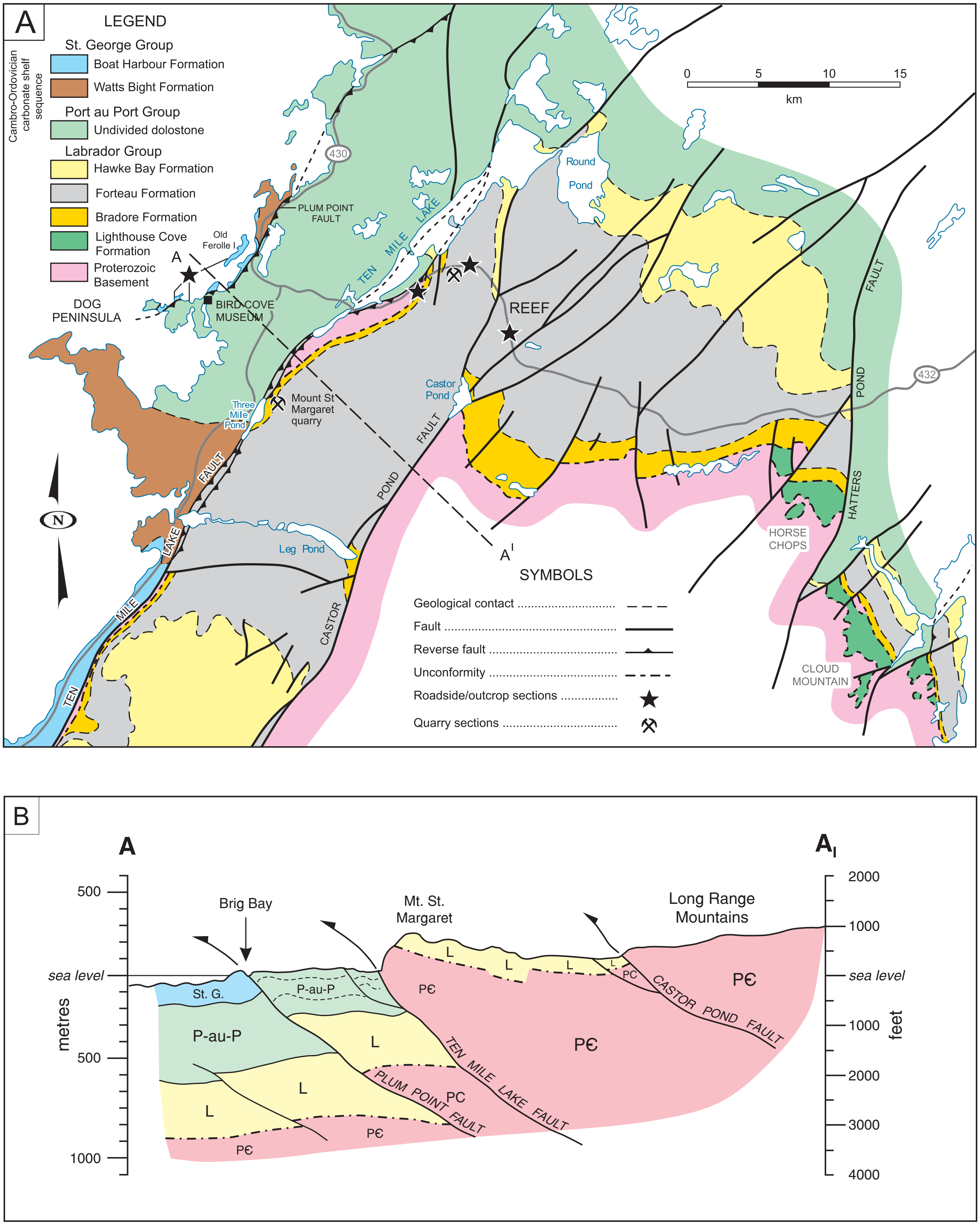

Figure 8. *A) Geology map of the Bird Cove to Canada Bay area showing location of sections and outcrops discussed in this article; B) A–A$_1$ - structural cross section; PЄ - Precambrian basement; L - Labrador Group; P-au-P - Port au Port Group; St. G. - St. George Group.*

in roadcuts above Ten Mile Lake where they include metamorphic gneisses and schists (Bostock *et al.*, 1983) and deformed tonalitic granites that date to 1610 Ma (Heaman *et al.*, 2004). The granitic rocks are grey to pink, crystalline in texture and have a faint structural fabric. White and grey-blue quartz, pink feldspar, and brown and white mica are common along with robust, hard crystals of dark-green pyroxene or hornblende. The granitoid rocks are intruded by dykes of black mafic rock of uncertain age but presumed to be Long Range dykes intruded during rifting (*see* below).

Large bodies of granitic rocks that outcrop along the northern edge of the Long Range Mountains intruded the gneisses at about 900 Ma (Bostock *et al.*, 1983). Grouped together in the **Grenville Province**, these crystalline rocks formed at great temperature and pressure during several episodes of orogeny deep in the Earth's crust (depths of more than 10 km, C.F. Gower, personal communication, 2010). Consequently as a result of convergence along a Pacific-type margin and eventually continent-to-continent collision of the Grenvillian southern edge of **Laurentia** with another ancient continental plate, **Amazonia** (as its name implies, now part of South America), the Grenville Province likely formed the keel of an ancient Himalayan-type alpine mountain belt. The amalgamation of the ancient Precambrian continents resulted in a single supercontinent, **Rodinia**, which was surrounded by a globally continuous ocean.

Recent paleomagnetic studies of time-equivalent Grenvillian rocks, elsewhere in Laurentia, show that the Grenville Province lay 30°N of the equator, 1300 Ma, drifting southward and crossing the equator to lie close to the south pole (>60°S) 300 million years later (Li *et al.*, 2008). By 900 Ma, the Grenville Province, including the Long Range Mountains area, lay landlocked within the southeast of the supercontinent, juxtaposed against Amazonia. Over the next 120 million years the supercontinent drifted north to lie above the equator, before drifting south of the equator again (Li *et al.*, 2008). By 800 Ma, although Rodinia began to break apart along its far western margin, the Grenville Province continued to be landlocked as Rodinia drifted from low to mid latitudes in the southern hemisphere; only 200 million years later did rifting reach the southeast part of Rodinia and begin to affect the rocks of the Great Northern Peninsula area.

The once deeply buried Grenvillian metamorphosed rocks in the Bird Cove area now lie unconformably below Cambrian sedimentary rocks (*see* below). This suggests that the Grenvillian alpine mountain belt was subject to massive erosion over long periods of time (hundreds of millions of years) to expose the Grenvillian rocks. The rocky mountainous landscape, barren of any terrestrial plants (they evolved later during the Silurian and Devonian; *see* website http://www.uni-muenster.de/GeoPalaeontologie/Palaeo/Palbot/seite3.html), was subject to the full force of physical and chemical erosion. At least some of this period would likely have included erosion during major global glaciations from 726 to 660 Ma and 655 to 635 Ma (*see* Hoffman and Li, 2009). Sturtian and Marinoan glacial deposits (as they are respectively known) are preserved in neighbouring parts of Laurentia such as northwest Scotland and North-East Greenland and are believed to signal a global snowball earth (*see* Hoffman *et al.*, 1998 and website http://www.snowballearth.org/). A younger Ediacaran glaciation, whose rocks are found on the Avalon Peninsula in eastern Newfoundland, dates to around 580 Ma. It also likely affected the study area as this was relatively soon after the first evidence of intracratonic rifting of the paleo-southeast of Rodinia (*see* below) and long before the rupturing of Laurentia from Amazonia around 550 Ma (Li *et al.*, 2008).

RIFTING: LONG RANGE DYKES AND THE LOWER LABRADOR GROUP

Late Proterozoic mafic dykes and sedimentary and volcanic rocks of the lower Labrador Group mark this event in northwestern Newfoundland (Figure 7). The first evidence of extensional rifting affecting rocks of the Great Northern Peninsula is a swarm of north- to northeast-trending, volcanic **Long Range dykes**. These black basaltic dykes, which date to 615–610 Ma (Figure 7; Kamo *et al.*, 1989; Gower *et al.*, 2001) cut the Grenvillian crystalline rocks along the eastern side of the Long Range Mountains. Locally, at Cloud Mountains and The Horse Chops above Canada Bay (*see* Figure 8A; Bostock *et al.*, 1983; Knight, 1987), the dykes appear to feed basaltic volcanic flows of the **Lighthouse Cove Formation**. Similar rocks are also known on Belle Isle and at Table Head and Henley Harbour in southeastern Labrador (Figure 7; Williams and Stevens, 1969; Strong and Williams, 1972; Bostock *et al.*, 1983; references in Gower *et al.*, 2001; Gower, 2010). The geological map at Canada Bay, Belle Isle and Table Head shows that the volcanic rocks lie upon eroded Grenvillian basement that, locally, was first veneered or buried by thin to thick deposits of red sandstone and conglomerate, known as the **Bateau Formation** (both rock units form the lower Labrador Group). The sedimentary rocks were locally tilted before the volcanic flows were extruded. As both the sedimentary and the overlying volcanic rocks are interpreted to be terrestrial and are restricted to narrow fault-bounded basins, a network of narrow rift valleys is envisaged to have characterized the area at about 610 Ma.

Evidence of rift rocks in the Brig Bay area is restricted to the mafic dykes at Ten Mile Lake. However, there is no evidence for rift sediments or volcanics along the west side of the Long Range Mountains either because there were none, or because they were subsequently eroded away before the Cambrian. Nonetheless, large faults, such as the Ten Mile Lake Fault and other faults in this area, were probably seismically active. An observer, at that time, standing in the vicinity of what is now Brig Bay would have both felt the earthquakes as rifting occurred and seen the ash clouds of distant volcanic eruptions near Canada Bay and Belle Isle.

By 550 Ma, Laurentia, which includes North America, Greenland, northwestern parts of the British Isles and perhaps parts of South America, was its own continent. A long narrow seaway, the beginning of the **Iapetus Ocean**, opened along the southern margin of Laurentia, to separate Laurentia from Amazonia; the latter remained with **Gondwana** (Li *et al.*, 2008). Paleomagnetic studies of the drifting continents suggest that the seaway (Iapetus Ocean, named for the father of Atlas in Greek mythology), opened to a major ocean by the time the earliest sedimentary rocks were laid down in the Bird Cove area at about 520 Ma *i.e.*, the late Early Cambrian (Winchester *et al.*, 2002; Li *et al.*, 2008). This means that for at least 80 million years following the rifting and volcanism discussed above, the area was subjected to continuous weathering and erosion that truncated the rocks of the Grenville Province and those of the lower Labrador Group.

CAMBRIAN UNCONFORMITY: AN ANCIENT LAND SURFACE

The basal Cambrian Unconformity that represents this ancient erosion surface is not exposed in the Bird Cove area. On the southern shore of Labrador and adjacent Québec (Gower *et al.*,

2001) and below a number of outliers of upper Labrador Group that form cuestas on the western edge of the Long Range Mountains inland of Hawkes Bay (Knight, 1991), the surface is essentially a gently undulating peneplain upon unweathered crystalline rocks. Only locally, is it ornamented with small, sediment-filled depressions as seen along the shore in Québec, immediately west of the provincial boundary. Patches of paleo-weathered and iron-enriched rocks are present below the unconformity near West St. Modeste in Labrador, and the western shore of Canada Bay (Bostock *et al.*, 1983; Knight, 1987, 1991; Gower *et al.*, 2001). Paleo-weathering of the top of the volcanic flows near Horse Chops and Cloud Mountain (Figure 8A) indicates that the late Proterozoic rift-bound rocks were also deeply weathered before the Early Cambrian sediments were laid down (Knight, 1987).

ROCKS OF THE TROPICAL MARGIN OF LAURENTIA IN NORTHWESTERN NEWFOUNDLAND

To discuss rocks of the early Paleozoic shelf of Laurentia there is no need to range outside the immediate study area to get a sense of what the earliest Cambrian world of this neighbourhood was like. Lying west and north of the Long Range Mountains, the coastal lowlands and carbonate barrens are underlain by rocks deposited in tropical, shallow-shelf seas along the southern edge of the Laurentian continent. The shelf flourished for more than 60 million years beginning about 520 million years ago, a time spanning parts of the Cambrian and Ordovician periods (Figure 2). Fossils preserved in these rocks show that some of the earliest forms of invertebrate animal life flourished in the Cambrian and Ordovician sea that lay about 20° south of the paleo-equator facing the Iapetus Ocean.

The Dog Peninsula and the area inland of Plum Point is host to three groups of sedimentary rock, the Cambrian Labrador and Port au Port groups and the Ordovician St. George Group (Figures 2 and 8A). Although each of these groups is separated from the other by a major fault, stratigraphic studies and mapping elsewhere on the Great Northern Peninsula indicate that the groups lie on the other in a continuous succession. As sedimentary rocks are laid down in sequential layers, the oldest layers buried by the youngest, the succession will be described from the oldest to youngest.

The succession is not only broken and disrupted by faults but is also incomplete because large parts of the succession have been eroded away. The mostly flat-lying to gently dipping sedimentary strata throughout the area is both advantageous and a hindrance to understanding the succession more fully. For example, the steep mountainous topography of Mount St. Margaret provides a vertical cross section through many tens of metres of strata representing many depositional events and a significant interval of geological time. Such a continuous rock sequence provides evidence to understand how the depositional setting on the ancient shelf during the Early Cambrian changed with time. In contrast, essentially flat-lying carbonate rocks exposed along the shoreline of Dog Peninsula represent only a few metres of succession and hence provide but a brief time window into rocks that were deposited as part of a much thicker and longer lived Late Cambrian to Early Ordovician succession. The shoreline however, commonly exposes large areas of bedding plane so that walking across these surfaces would be analogous to walking about on the ancient seafloor.

LABRADOR GROUP OF MOUNT ST. MARGARET AND THE TEN MILE LAKE AREA

The oldest sedimentary rocks in the Bird Cove area belong to the upper Labrador Group, so named by Schuchert and Dunbar (1934) for the well-exposed sedimentary rocks of southern Labrador. These fossiliferous rocks consist of sandstone, shale and limestone. The group consists of the Bradore, Forteau and Hawke Bay formations. Several quarries and roadcuts along Route 432, east of Ten Mile Lake and at MSM near Three Mile Pond, provide excellent exposures of the rocks of the Forteau Formation. The basal Bradore Formation is generally poorly exposed but can be seen in some outcrops on the access road to the MSM quarry. The uppermost rocks of the group are not described because, although present in the area, they cap the Highlands of St John (Figure 3) and are inaccessible. They are well exposed along the south shore of Hawkes Bay (Knight, 1991).

Bradore Formation

The Bradore Formation sits unconformably upon Grenvillian basement and is known to be at least 80 to 150 m thick (Schuchert and Dunbar, 1934; Cumming and James *in* Bostock *et al.*, 1983; Knight, 1991). It consists of red, brown to purplish, coarse- to fine-grained, sub-arkosic sandstone, rich in pink feldspars and white and blue-grey quartz. Close to the base of the formation, the sandstone is pebbly (mostly quartz, lesser feldspar, rare granite and some red mudstone pebbles) and rich in black laminae of heavy minerals such as magnetite and haematite. Crossbedding is common in lower strata but upward is largely obliterated by widespread development of burrows. The long narrow, vertical tubes known as *Skolithos linearis* Haldeman, 1840 are believed to be constructed by polychaete worms or worm-like creatures (perhaps like modern phoronids horseshoe worms) that filtered suspended nutrients from the seawater. These same burrows are found in lithologically similar time-equivalent rocks of the Eriboll Formation near Durness and Loch Eriboll in northwest Scotland where they are known as 'The Piperock' (McKie, 1990).

The few isolated outcrops of the formation seen along the east side of the access road past Three Mile Pond Park to the MSM quarry show some of these features, but to see this formation at its best, it is recommended that the southern shore of Labrador be visited. There, the full vertical section through the formation is traceable over about 35 km of strike and along shoreline cliffs, hillsides, roadcuts and quarries of southern Labrador from Blanc Sablon to Diablo Bay. The sandstones were believed to pass from basal largely non-marine, river-lain deposits of a southeastward-flowing, braided fluvial system up into nearshore sandstones that were extensively burrowed in a sheltered estuarine setting below uppermost sediments, deposited in a tidal inlet-sand bar complex (Hiscott *et al.*, 1984; Knight, 1991). Recent study of the Labrador sections, however, suggests that the formation represents a gradually deepening shallow-marine shoreline sand sheet, deposited on a storm-influenced, wave-swept shelf (Long and Yip, 2009).

Forteau Formation

The Forteau Formation throughout northwestern Newfoundland comprises a three-part stratigraphy, namely a basal carbonate known as the **Devils Cove limestone**, a middle shale-dominated interval that coarsens in its upper half and an upper mixed carbonate and clastic interval which on the Great Northern Peninsula, includes **archaeocyathid** (pronounced ark-e-o-sigh-ath-id) reefs (Knight, 1991). The succession at MSM and Route 432 east of Ten Mile Lake belongs to the lower 2 intervals and the basal part of the upper one (Figure 9). The shale of this succession is potash rich (up to 10%, Swett and Smit, 1972). Because the easily weathered shale has been used extensively over the last 50 years or more to construct the local road-beds, vegetable gardens have long been set in the shaly soils of abandoned quarries and along the roadside.

The Devils Cove limestone, named for a cove in Chimney Arm, Canada Bay, is no more than 3 to 4 m thick, in broken outcrop, at the entrance to MSM quarry, and in a roadside outcrop along the East Castor Pond (ECP) resource road. At the quarry, it overlies very coarse to granular, pebbly sandstone, whereas at the resource road it is intercalated with red Bradore sandstone. At the quarry, the carbonate consists of pink and white, crystalline dolostone overlain by fossil-rich, flaggy-bedded grainy limestone. Crinoid debris dominates the grainy carbonates in which are scattered archaeocyathids, trilobites and calcareous articulate and phosphatic inarticulate brachiopods. Similar, vuggy, crystalline dolostone and limestone, hosting archaeocyathid debris, marks the base of the unit at ECP below very coarse grained red sandstone. Above the sandstone, red and grey, bioturbated, fossiliferous, fine-grained limestone is stylobedded and nodular with red and grey shale partings. The limestone is rich in green glauconite, a mineral indicative of shelf settings; fossils include articulate and inarticulate brachiopods, trilobites and crinoid debris.

The middle shale, 60 m thick, is exposed in the various quarry faces at MSM quarry (Figures 10 and 11) and in part in shale quarries along Route 432. The dark-grey shale is not uniform up through the member. It begins with a poorly exposed 20-m-thick section, in the floor of the outer workings of the old quarry (Figure 10A and B). The first few metres consist of burrowed silty mudstone and shale below grey shale and scattered carbonate nodules. The grey shale is intercalated about every 1 m with 10- to 20-cm-thick interbeds of fossiliferous, lumpy and nodular, grey, fine-grained, micaceous limestone, rich in comminuted phosphatic shell debris and apparently intensely churned by coarse bioturbation, likely *Thalassinoides*. Pyrite is common in the shale, in burrows and may encrust and replace fossils. The fossils, which are generally thin shelled, include phosphatic inarticulate brachiopods such as *Obolella* and *Paterina* (C. Skovsted, personal communication, 2007), the cone-shaped agmatan *Salterella*, hyolithids, olenellid trilobites and rare crinoids (Figure 11).

The middle 20 m occurs at the older quarry floor and faces at MSM and in the Route 432 quarry (Figure 10). Except for a rare siltstone or limestone ribbon and a few horizons of scattered yellow-weathering, golf-ball-sized carbonate nodules (Figure 10C), it is dominated by shale. The nodules in all intervals can range from structureless to those that preserve lamination, crosslamination, burrows and fossils (Figure 11C and D)[1]. The shale has good fissility and appears featureless at first sight although bioturbation becomes clearly visible on some bedding plane surfaces

[1] The nodules provide a useful lesson in the effect of compaction of mud to shale and mudstone. During this process, the mud may compact to 65 to 80% of its original thickness by loss of water and flattening of clays and other minute mineral grains into paper thin layers of fissile shale. Because the carbonate nodules formed as hard mineralized concretions that nucleated and replaced the sediment soon after it was deposited, they preserve uncompacted sedimentary structures in the mud as well as the skeletal shape of fossils. Trilobites and hyolithids commonly recovered from the shale in the quarries are usually flattened to paper thin imprints/moulds on the shale fissility (*see* Figure11A and B). However, uncompacted fossils found as part of the nodules show that the trilobites and hyolithids are robust, shapely carapaces and cones (*see* Figure 11C and D); in particular a nodule preserving an *Elliptocephala* thorax and pygidium (tail) shows strong tail ribbing and scaly patterning that gives the impression of a robust, well-armoured predator.

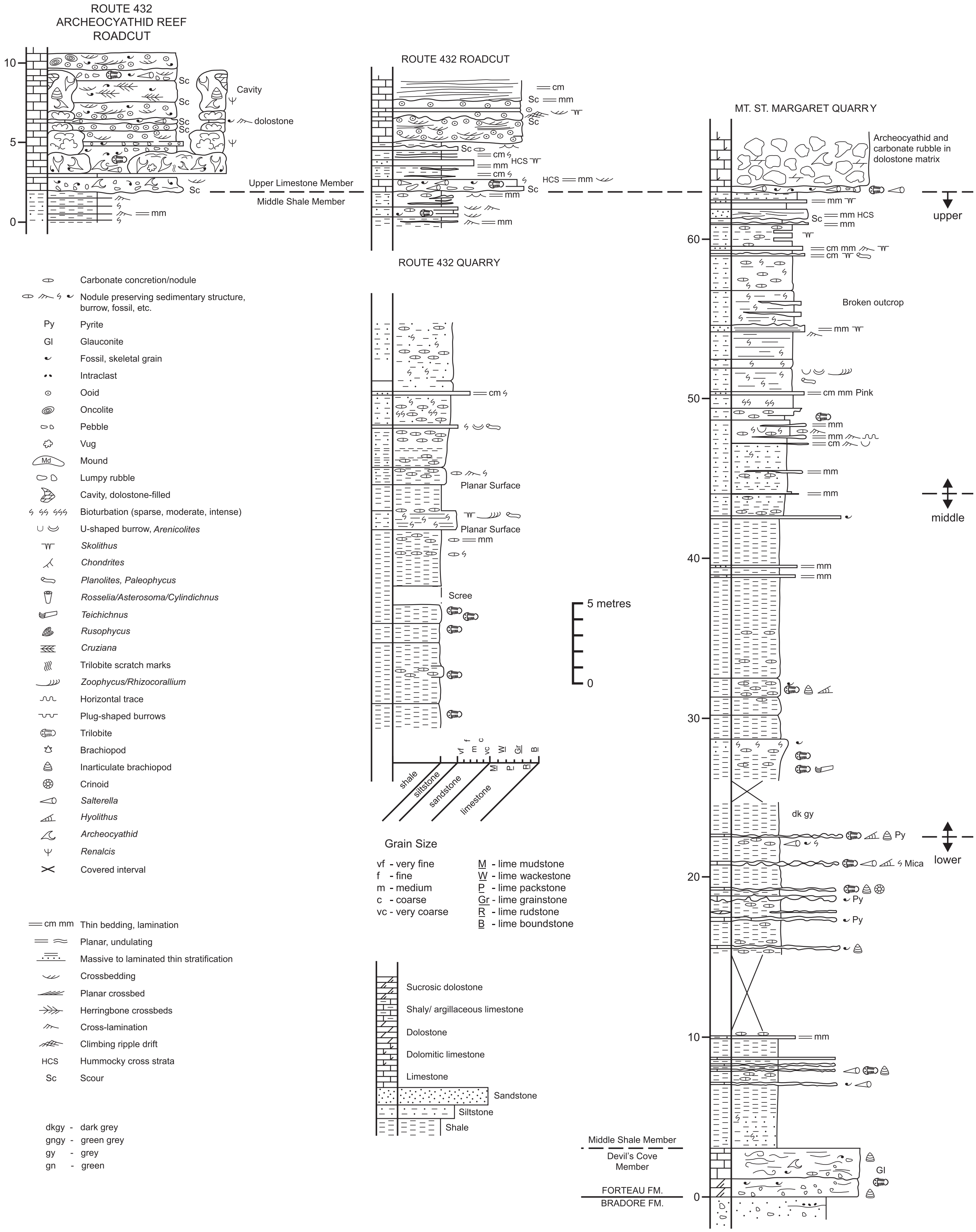

Figure 9. *Stratigraphic sections measured through the Forteau Formation at Mount St. Margaret and roadcuts on Route 432.*

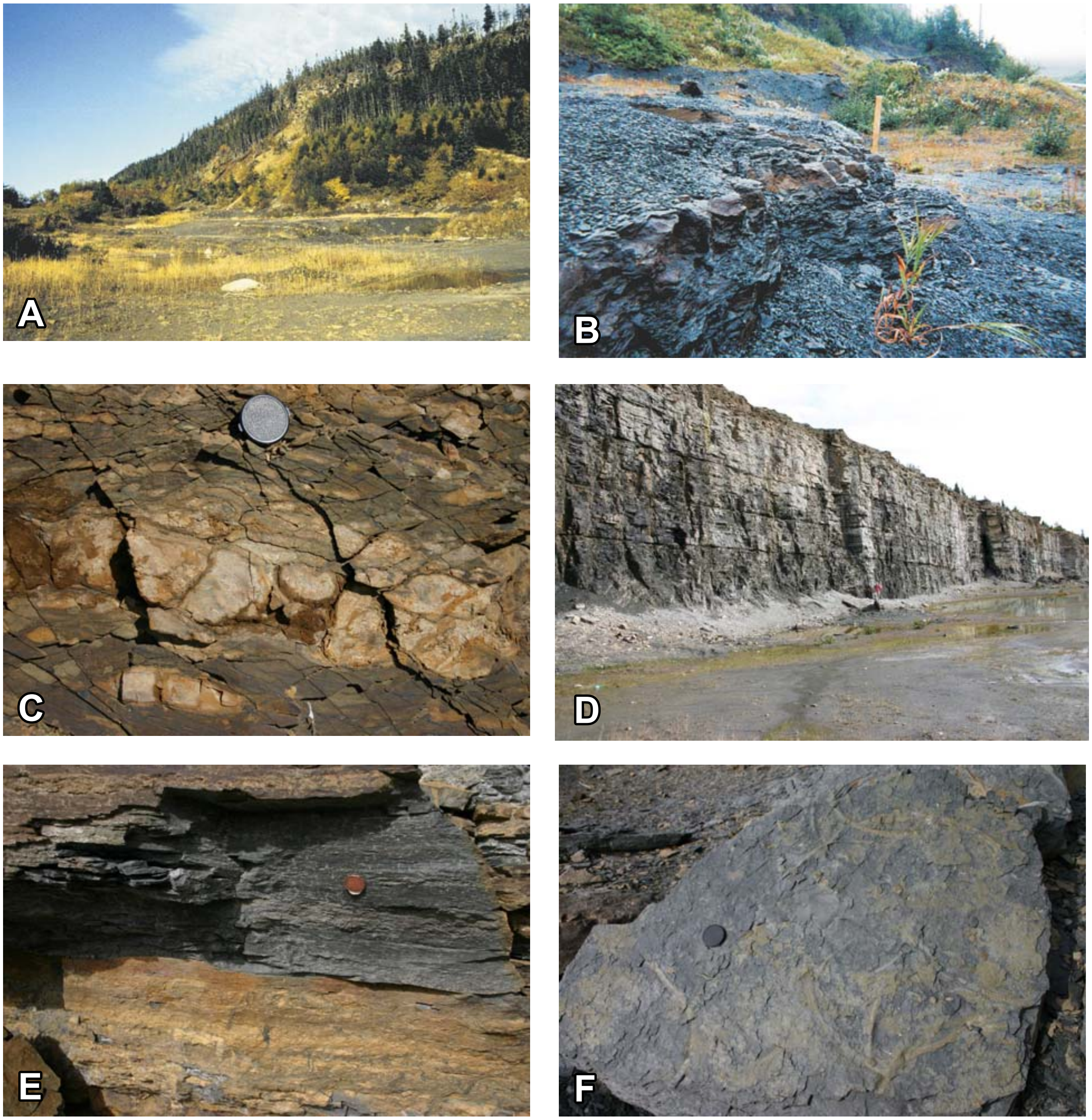

Figure 10. *Forteau Formation at Mount St. Margaret. A) The older quarry in lower part of the shale member. Trilobites are common in the low benches below the trees; B) A fossil-rich limestone bed in the shale; both shale and limestone are fossiliferous; C) Carbonate nodules in the shale; D) South quarry in upper part of middle shale member; E) Intensely bioturbated siltstone and shaly mudstone in upper shale; F) Large burrows on the base of a large block in the south quarry; G)* Skolithos *tubes vertically penetrating a laminated fine-grained sandstone, south quarry.*

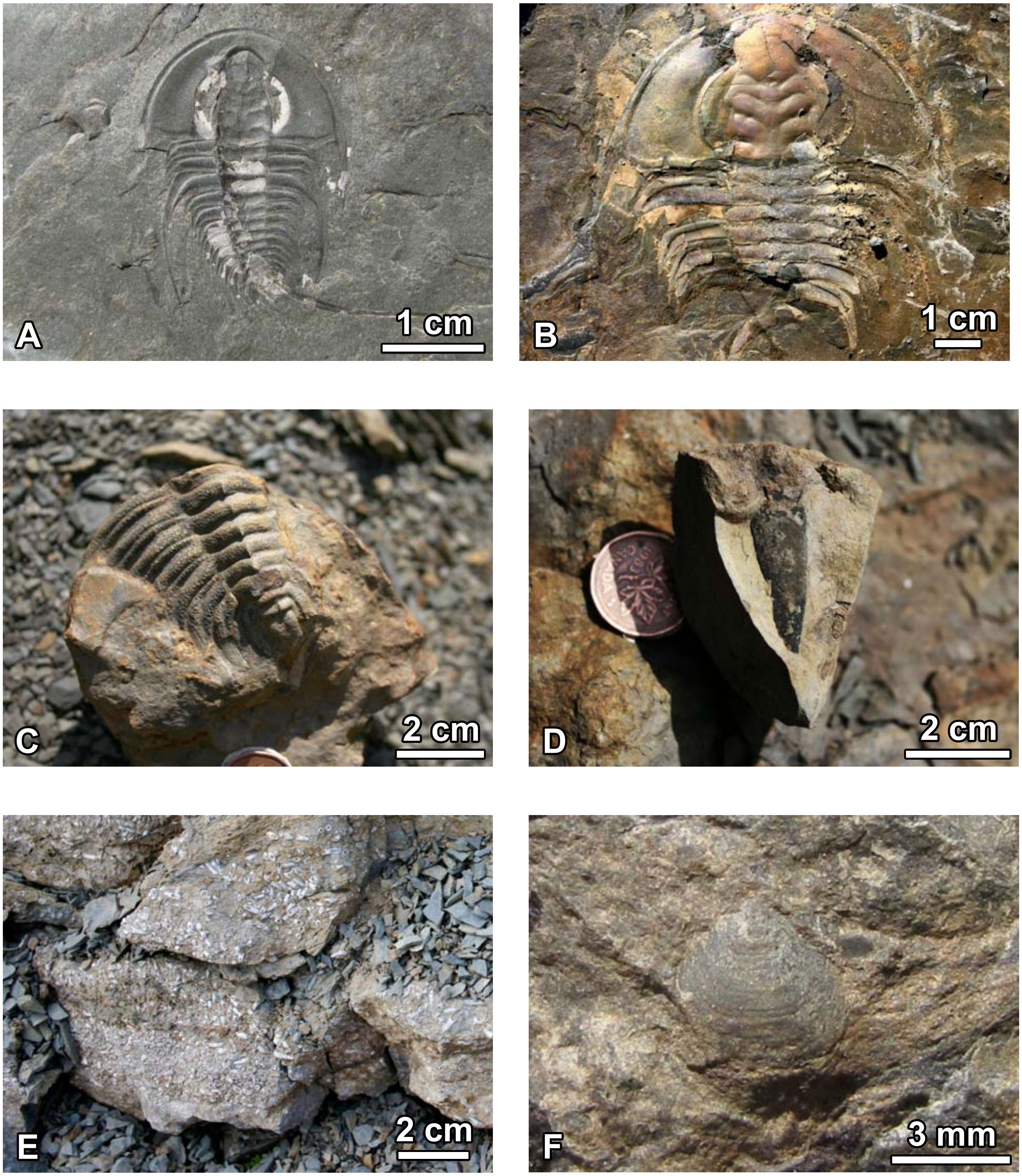

Figure 11. *Shelly fossils from the Forteau Formation. A and B)* Olenellus transitans *(Walcott, 1910); C) Possible thorax and pygidium of* Elliptocephala logani *(Walcott, 1910); D)* Hyolithus *sp. indet.; E) Clustered small cones of* Salterella*; F) Pedicle valve, inarticulate brachiopod,* Obolella chromatica *Billings, 1861. (Note: Figure 11A to D and F from middle shale; Figure 11E from upper limestone east of Hawke Bay. Fossils in Figure 11A to D and F housed at The Rooms, Provincial Archive and Museum, St. John's.)*

under natural light. The latter include tubular burrows, and possible *Teichichnus*. Disarticulated and complete olenellid trilobites (Figure 11A and B) grace the fissility of the middle shale interval and are especially well preserved in the lower benches of the shale quarry on Route 432. The most common trilobites are *Olenellus transitans* (Walcott, 1910)[2], a trilobite with a generally smooth carapace, fine terrace lines and impressive tail and thoracic spines (Figure 11A), and *Elliptocephala logani* (Walcott, 1910), a form with a distinctive reticulate surface ornamentation (scaly looking patterning) on the carapace (prosopon; Figure 11C). Other trilobites known from the shale include *Elliptocephala* sp. undet. and the eodiscid *Calodiscus lobatus* (Hall, 1847). Like modern day crabs and lobsters, trilobites possessed a hard exoskeleton and grew by periodically shedding their too small exoskeleton (moulting) and growing a new larger shell. Consequently, trilobites in the shale range from disarticulated parts of various sizes and growth stages to complete trilobites that represent individuals. Trilobites lived by scavenging, preying on other smaller organisms on, and above, the seafloor and by grazing on microbial algal mats.

The upper 20 m of the shale member is exposed in the face of the newest quarry to the south of the old workings at MSM and in the inner pit of the quarry on Route 432. It consists of well stratified decimetre-thick beds of intensely bioturbated muddy siltstone and lesser shale in which most depositional structure is destroyed by the bioturbation (Figure 10E); lamination and cross-lamination are locally preserved and limestone nodules also occur. The member is, however, increasingly punctuated upward by distinct, resistant beds and lenses of coarse-grained siltstone to very fine-grained sandstone and rare limestone. They display lamination, crosslamination and burrows. Some sandstone beds are 50 cm thick near the top of the member; they are characterized by scoured bases, internal lamination, hummocky cross-stratification (HCS) and locally ripple-marked tops (Figure 10G). An abundant assemblage of burrow and bedding plane trace fossils is dominated by a distinctive, vertical to inclined, concentrically lined tubular burrow, likely *Cylindrichnus* and/or *Rossellia*, but also includes simple tubes such as *Planolites* and more complex traces such as *Chondrites*, *Teichichnus*, and rare *Rhizocorallium* or possible *Zoophycus*. *Skolithos* and U-shaped tubes (*Arenicolites*) penetrate the laminated and HCS sandstone and siltstone beds in the upper half of the interval (Figure 10G). The trilobite traces *Cruziana*, *Rusophycus* and scratch marks occur on the bases of the coarser beds. Fossils (trilobites, archaeocyathids and molluscs) have only been recovered from limestone beds intercalated just below the top of the member. Large blocks scattered on the floor of the quarry and in the rubble of an unexcavated shot blast close to the road as it ramps up the hill provide excellent opportunities to see the variety of bioturbation (Figure 10F) and other sedimentary structures, such as ripple marks.

The basal strata of the upper member of the Forteau Formation occur in the treed hillside above the MSM quarry and in equivalent limestone strata of a number of roadcuts and small quarries along Route 432, east of Ten Mile Lake. A 3 m carbonate conglomerate resting erosively upon a skeletal grainstone rich in trilobites (*Bonnia* and *Elliptocephala*) and *Salterella* marks the base of the member at MSM. The carbonate conglomerate of limestone clasts and rare archaeocyathids has a yellow, argillaceous dolostone matrix. Above this, the succession, although poorly exposed, consists of black, grey and red grainstone, a limestone formed of sand-sized carbonate grains, notably oolite, oncolite, and various skeletal grains. Associated with these grainstones at MSM and at one roadcut on Route 432 are reefs built primarily of archaeocyathids, an Early Cambrian

[2] This has also been identified as *Olenellus thompsoni* (Hall, 1859).

reef-building calcareous organism affiliated with the sponge family. The archaeocyathan mounds and reefs thrived widely in the Early Cambrian across many of the scattered continents of the time (James and Debrenne, 1981). At MSM, the reefs of red and grey limestone are best seen along the gravel road to the telecommunications tower above and east of the quarry. There, the planed-off outcrops show reefs that range up to 14 m wide surrounded and separated by up to 12 m of inter-reef grainstone including oolitic grainstone.

The reefs and associated grainstone are better exposed in low cliffs either side of a roadcut at the brow of a hill on Route 432 (Figure 12), 21 km from the junction with Route 430. Several reefs occur at this locality, each about 6 m wide and high, part of a series of east-trending, parallel reef walls, each separated by inter-reef channels of different widths (Figure 12A and B). A thin layer of ripple, crosslaminated dolostone midway through the reef bodies can be traced to a bed in the inter-reef channel grainstones showing that the reefs likely grew in stages and were roughly in sync with their channel fills.

The reefs are built of an open framework of densely packed to open clusters of large sheets, cups, sticks and branching bushes of a variety of archaeocyathids (Figure 12C). Only three archaeocyathans are known in this part of the Forteau Formation, namely *Archeocyathus atlanticus* Billings, 1861, *Metaldetes profundus* (Billings, 1861), and *Retilamina amourensis* Debrenne (Debrenne and James, 1981). Small, dark-grey, bush-like structures believed to be the calcimicrobes *Epiphyton* and *Renalcis* also occur as a minor constituent of the reefs. They flourished in irregular open spaces in the reefs, attached to and sheltered below the archeocyathans (James and Kobluk, 1978; Kobluk and James, 1979). Irregular patches of fossiliferous grainstone, and grey and red structureless, fine-grained limestone and laminated dolostone scattered throughout the reefs are interpreted as open space and cavities in the reefs infilled by sediment. The latter was produced within the reef or washed in by currents that surged back and forth around and through the reefs as they grew. Detailed studies of reefs exposed in southern Labrador indicate that cavities were inhabited by numerous, small, encrusting organisms such as algae and foraminifera; were cemented by fine-grained micritic limestone and layered fibrous and blocky calcite cements; and were partially or completely infilled by fine-grained mud, known as geopetal sediment[3] (James and Kobluk, 1978; Kobluk and James, 1979). Organisms also bored the reefs. Similar structures are present in the reefs on Route 432 but no detailed study has been completed to compare them to the classic Labrador examples.

Deposits of the inter-reef channels (Figures 9 and 12A and D) include skeletal grainstone, intraclastic-skeletal grainstone and oölitic grainstone[4]. The intraclastic–skeletal grainstone is generally coarser grained than the oolitic grainstone and include tabular rip-up clasts of limestone up to 15 cm long; erosional surfaces are frequent. The skeletal grainstone is dominated by crinoidal debris and the millimetre-sized cone-shaped agmata *Salterella* but also includes trilobite debris (*Bonnia*,

[3] Geopetal sediment – fine sediment that filters down through the quiet (although still gently moving/circulating) waters filling interconnected reef cavities to settle on the floors of cavities and tops of organism as fine laminated sediment. Hence it is actively accumulating at the same time that cements are precipitating on the roofs and walls of cavities.

[4] Oolites are sand-sized grains of limestone formed when layers of needle-like crystals of calcium carbonate (usually aragonite) are precipitated systematically around a nucleus, as the grains are suspended above and rolled about on the shallow seafloor by currents. Fossil fragments, microchips of eroded limestone, phosphate or quartz sand may form the nucleus of the oolites. Well-known examples of modern day oolite sands occur along the southwestern outer edge of the Bahama Banks where tidal current and wave-driven, nutrient-rich, ocean water interacts with the strongly saline and warm, bi-carbonate rich waters of the shallow banks. The vast active oolite sand complexes form in water as shallow as 3 m and are characterized by large ripple marks, sand ridges, dunes and bars (Ball, 1967; Tucker and Wright, 1990; Rankey *et al.*, 2006).

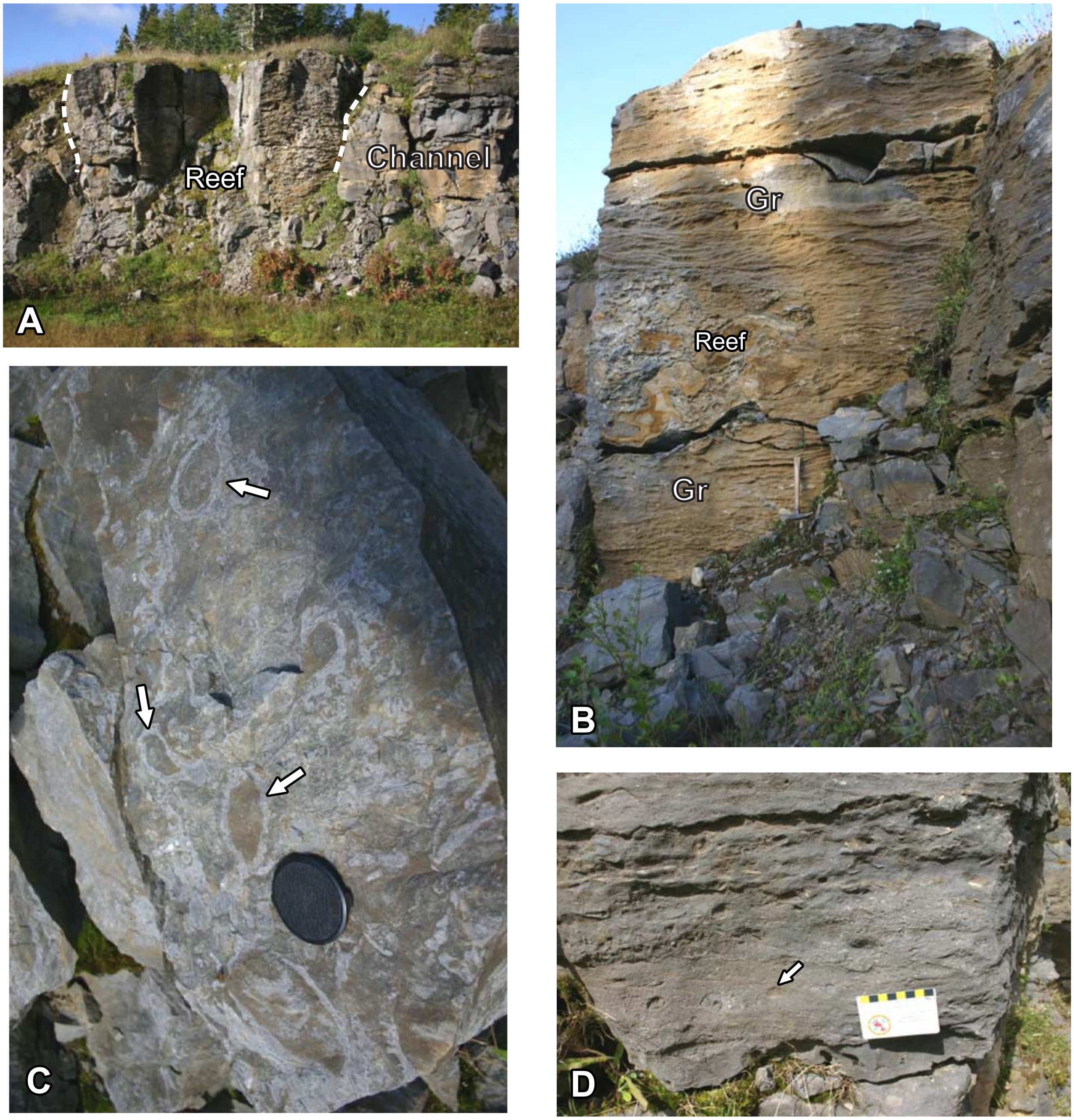

Figure 12. *Archaeocyathid reefs and grainstone-filled channels, upper limestone member, Forteau Formation, roadcut, Route 432. A) A reef buttress abutted by bedded grainstone; B) Edge of a reef enclosed by current-bedded and crossbedded grainstone (Gr) of inter-reef channel; C) Block displaying cross-sections of archaeocyathid cups and sticks; D) Black oolitic grainstone with some skeletal debris, a few limestone pebbles and a rare oncolite (arrow).*

Olenellus and *Elliptocephala*). Although archaeocyathid intraclasts occur in the grainstone at the base of the reefs, intraclasts higher in the channel appear to be derived by erosion of lithified limestone from the seafloor away from the reefs. The oolite grainstone occurs as distinct beds of medium-grained and fine-grained oolite. Rare oncolites[5] also occur. The grainstones are trough crossbedded[6] and show that currents were dominantly westward directed although some were eastward. The opposing two-way directions imply tidal currents likely operated across the reefs.

The Route 432 reefs can be traced southwest to ridges along the East Castor Pond resource road to MSM. They are also known in the Highlands of St. John (Schuchert and Dunbar, 1934) and as a single bed along Torrent River just inland of the fish ladder at Hawkes Bay (Knight, 1991). The reef facies[7] has not been mapped east of this belt.

The top of the reefs and inter-channel grainstones are truncated and overlain by a bed of crossbedded, skeletal-oolitic-intraclastic grainstone, 1.5 m thick. No reefal material has so far been found in this bed.

Depositional Environment of the Upper Labrador Group

The upper Labrador Group succession of the Bird Cove area, reflects the interplay of rising global sea level and sedimentation along the margin of Laurentia throughout the late Early Cambrian. At this time, the shelf was a relatively narrow, gradually deepening ramp with land to the present day northwest and ocean to the southeast (James *et al.*, 1989). Weathering of the Laurentian continental craton provided abundant terrigenous sediment[8] that was carried by braided rivers across a likely expansive floodplain to the shoreline of the Iapetus Ocean. There, storm waves sorted sands into a broad, shoreline-hugging sand sheet characterized by rapidly shifting sand bodies (Long and Yip, 2009). Filter-feeding worms with the ability to respond to the rapidly deposited, shifting sand bodies and the harsh storm-dominated conditions were the only organisms to thrive in this setting.

The coastal floodplain and shoreline moved farther into the continental interior as continuing late Early Cambrian sea-level rise (transgression) drowned the margin. Consequently, terrigenous sediment was trapped close to the shore and little clastic detritus reached the open seas of western Newfoundland for a short period of time. This likely resulted in clear seas at this stage of the transgression, and in this setting, calcareous skeletal organisms including abundant crinoids, trilobites, brachiopods and scattered archaeocyathids, flourished to form the Devils Cove limestone at

[5] Oncolites – ball-like internally laminated algal structures related to stromatolites that were moved freely about on the seafloor by currents as the algal balls grew.

[6] The crossbedded grainstone (*see* Figure 14) preserves the internal layered structure of sand waves and large ripples, which were fashioned by strong, perhaps tidal, currents, as they swept through the channels between the reefs. Each sloping sediment layer in the crossbeds represents a depositional event as ripples and dunes migrate in a down-current direction across the seafloor.

[7] Facies -term used to define specific sediments/sedimentary rocks with a range of distinct characteristics (such as mineralogy, sedimentary structures, fauna, burrowers) that reflect specific depositional and environmental conditions.

[8] It is worth remembering that there were no land plants to cover the Cambrian continents so that weathered rocks were easily eroded by all available physical processes and moved into an adjacent river system for transport to the ocean. Storm-dominated seas would also easily rework deeply weathered crystalline Grenvillian basement into a thick, widespread basal marine sand deposit as sea level rose (transgression) and drowned the ancient continental margin.

the base of the Forteau Formation. The crinoids, which are better thought of as upside down echinoids that have long tentacles and are fixed by long stalks to the seabed, filtered suspended particles from the water column. Often called sea lilies, they likely formed meadows across the shelf in the Bird Cove area; trilobites and other seabed-dwelling shelly organisms sought shelter in the meadows. Small archaeocyathid reefs are only known in the unit in Labrador. However, their presence as solitary fossils in the Devils Cove limestone in northwest Newfoundland suggest they were either transported from the nearest reef to the sea lily meadows by storms or they struggled but ultimately failed to get established in this part of northwest Newfoundland.

Continued, likely rapid, transgression overdeepened the shelf as it reached its maximum. Consequently, environmental conditions no longer favoured carbonate production in the MSM area even though numerous archaeocyathid reefs graced the shelf 60 km to the north on the Labrador coast. The archaeocyathid reefs in Labrador are interpreted as patch reefs that preferred moving but not turbulent water and hence thrived in relatively sheltered, shallow-shelf settings surrounded by muddy and silty shelf sediments (James and Kobluk, 1978). Fine-grained sediment, principally mud, was carried to the deeper shelf in northwest Newfoundland likely by muddy bottom flows from distant river mouths and by storms stirring up sediment in shallower parts of the near-shore shelf and redepositing it farther out on the shelf. The mud was redeposited below storm wave-base in quiet deeper waters to form the thick shale interval. Although the dark-grey Forteau shale contains some pyrite indicative of generally low oxygen levels in the sediment themselves, there is much evidence that the shelf bottom was still hospitable for bottom-dwelling organisms. Abundant trilobites and hyolithids in the shale, *Salterella*, trilobites, hyolithids, brachiopods, small molluscs and crinoids in the few limestone interbeds, and evidence of burrowers in the mud suggest that the seafloor was not a dead zone but a thriving eco-system of small shelly organisms, insediment burrowers and crawling and swimming arthropods. Beds of fine-grained limestone suggest transport of lime mud to the shelf from shallower settings or intervals when the bottom conditions allowed limestone to form on the shelf.

Sedimentary structures, such as lamination and crosslamination, preserved in many carbonate nodules in the shale provide strong evidence that the mud settled from suspension through the water column and was also reworked and transported along the seabed by weakish currents; burrows in some nodules show there were also a variety of mud-inhabiting organisms. The shelf was likely at its deepest during deposition of the middle part of the shale.

Once above this middle shale interval, the clastic succession, although still relatively fine grained, coarsens upward. This suggests that the shelf became shallower as sediments built seaward (*i.e.*, prograded) as frequent graded, muddy siltstone–shale beds and infrequent but distinct coarser beds of quartz sandstone were deposited. The rhythmic repetition of the graded rock types suggests storms repeatedly moved sediment from the near shore to the outer shelf where, once deposited, it was colonized by a vigorous burrowing, largely soft-bodied, animal fauna. The distinct sandstone beds indicate that shoreline sands were at times swept out onto the shelf. The presence of U-shaped burrows and *Skolithos* in the sands suggests that they were colonized by worms that included suspension feeders.

Cutoff of clastic sediment to the shelf at the end of this time reduced turbidity of the sea and allowed the re-establishment of a carbonate shelf in the upper limestone member. The complex of archaeocyathid reefs and channel-bound, crossbedded skeletal, intraclastic and oolitic grainstone, seen on Route 432, MSM and Highlands of St. John suggests a shallow-shelf setting strongly influenced by currents as modern oolitic lime sands are generally deposited in a few metres of water, *i.e.*, close to sea level. The east–west-oriented reef-channel complex suggests that the high-energy environment was shaped by local and perhaps regional hydrodynamic forces of prevailing weather and currents on the Forteau shelf during the late Early Cambrian. Also, as the reefs have not been mapped east of the Route 432 outcrop and in other eastern sections of the Northern Peninsula, it is possible that a roughly northeast- to north-trending wave resistant barrier occurred on the Forteau shelf in this northwestern part of the Northern Peninsula, east of which the shelf was dominated by mud deposition as it progressively deepened (Figure 13). Thick bodies of oolitic limestone and sandstone in the Hawke Bay area (Knight, 1991) would also form part of this barrier complex, suggesting that reefs were not ubiquitous everywhere along the barrier.

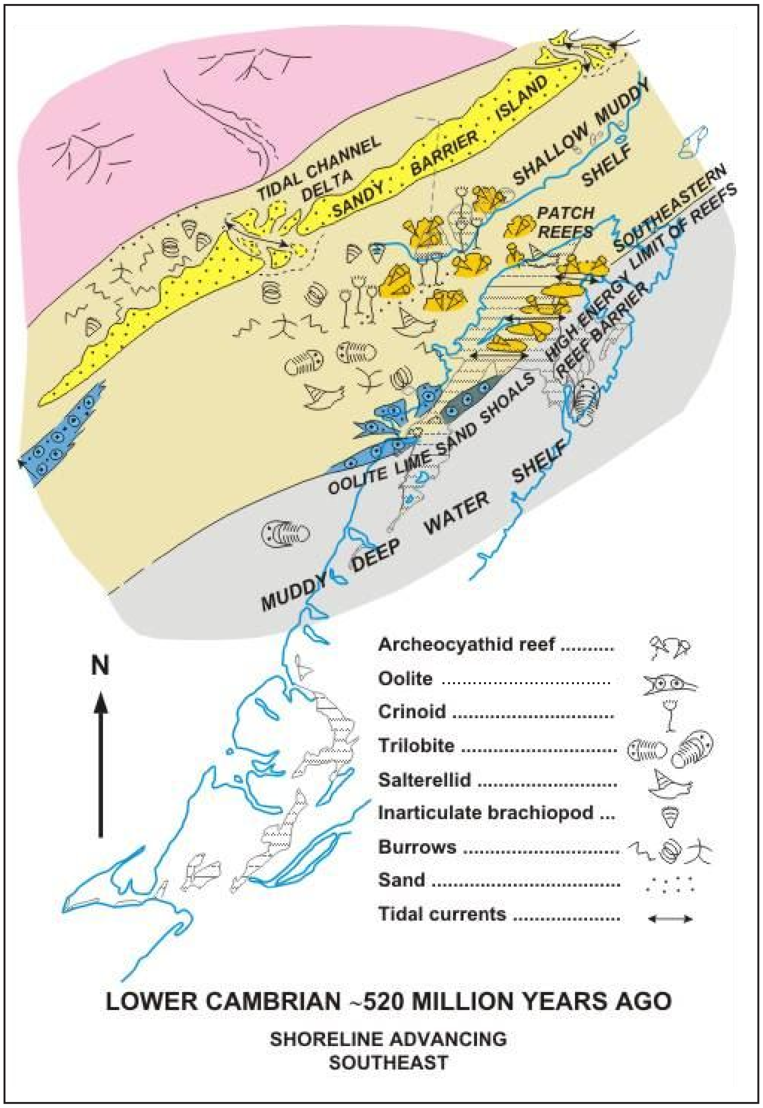

Figure 13. *A simplified map of the Early Cambrian shelf in the area of the Great Northern Peninsula and Labrador about 520 million years ago.*

LATE CAMBRIAN PORT AU PORT GROUP, DOG PENINSULA: DEPOSITS OF A SHALLOW, SALTY SEA

Late Cambrian, light grey dolostone of the upper Port au Port Group underlies much of the coastal lowlands west of the Ten Mile Lake Fault (Figure 8A). The group is generally poorly exposed and because of this, it is difficult to subdivide it into formations. In the Bird Cove area, however, chert is widespread suggesting that the dolostone belongs to the youngest formation in the group, the **Berry Head Formation**. This is supported by outcrops along Route 432, west of Ten Mile Lake and on Mutton Island in Ste. Genevieve Bay that host the small, low-spiral coiled gastropod, *Sinuopea* (Boyce, 1979; Snow and Knight, 1979; Knight, 1986; Rohr *et al.*, 2000). Large, chert-rich microbial mounds of dark-grey, sucrosic dolostone are interbedded with the light-grey dolostone on many of the islands in Ste. Genevieve Bay. The mounds are mostly characterized by columnar stromatolite structure. These mound-rich dolostones are a distinct marker at the base of the Berry Head Formation elsewhere in western Newfoundland, best seen on the point near Anchor Point, just north of the Bird Cove area.

Late Cambrian Tidal Flats and Algal Mounds – Meany's Point and Long Bottom

The coastal exposures of the group at Meany's Point and Long Bottom near Bird Cove provide an opportunity to examine a short section of strata (Figure 14) deposited in a lagoonal and tidal flat setting. The almost flat-lying bedding surfaces that can be examined over relatively long distances, expose broad expanses of an ancient seafloor. Fossils are rare in the dolostone; only beach rubble at the point has yielded small black, inarticulate brachiopod shells. The succession, which is no more than 6 m thick, consists of three rock types: burrowed dolostone; mound dolostone; and thinly bedded, sometimes shaly, dolostone (Figure 15). The three rock types are viewed as metre-scale, upward-shallowing cycles of burrowed dolostone, with or without mounds, overlain by thin-bedded dolostone. However, as thin beds of burrowed dolostone intercalate within the thin-bedded facies, the cyclicity may be more conjecture than real!

The burrowed, fine-grained, grey dolostone is characterized by flaggy thin bedding, simple tubular burrows, and locally wrinkled to pin cushion-like irregularities on bedding surfaces suggesting microbial cyanobacterial algal mat (Figure 15A, B and C). The burrowed dolostone is closely associated with the mounds (Figure 15D); the latter overlie and are draped by the burrowed dolostone. The robust, steep-sided bun-shaped mounds sit on flat, gently rippled and burrowed dolostone. The slightly oval mounds, each about a metre in diameter and rarely more than a metre high, also coalesce into large complexes up to 8 to 10 m wide with a broad, flattish but pillowed, top resembling a batch of dinner rolls. Smooth shallow furrows occur between the mounds of the pillowed top and continue down the steep sides of some complexes. Flat disc-like pebbles are locally wedged together in the narrow spaces between tops of mounds. Internal structure of the mounds is unknown but by comparison to similar mounds elsewhere along the Great Northern Peninsula shoreline, *e.g.*, Seal Ledges at Flowers Cove, they are probably thrombolitic mounds.

Elongate, although somewhat sinuous, mound-free areas, up to 10 m wide, wander between the mound complexes. These areas roughly trend due south, and host patches of small, straight-

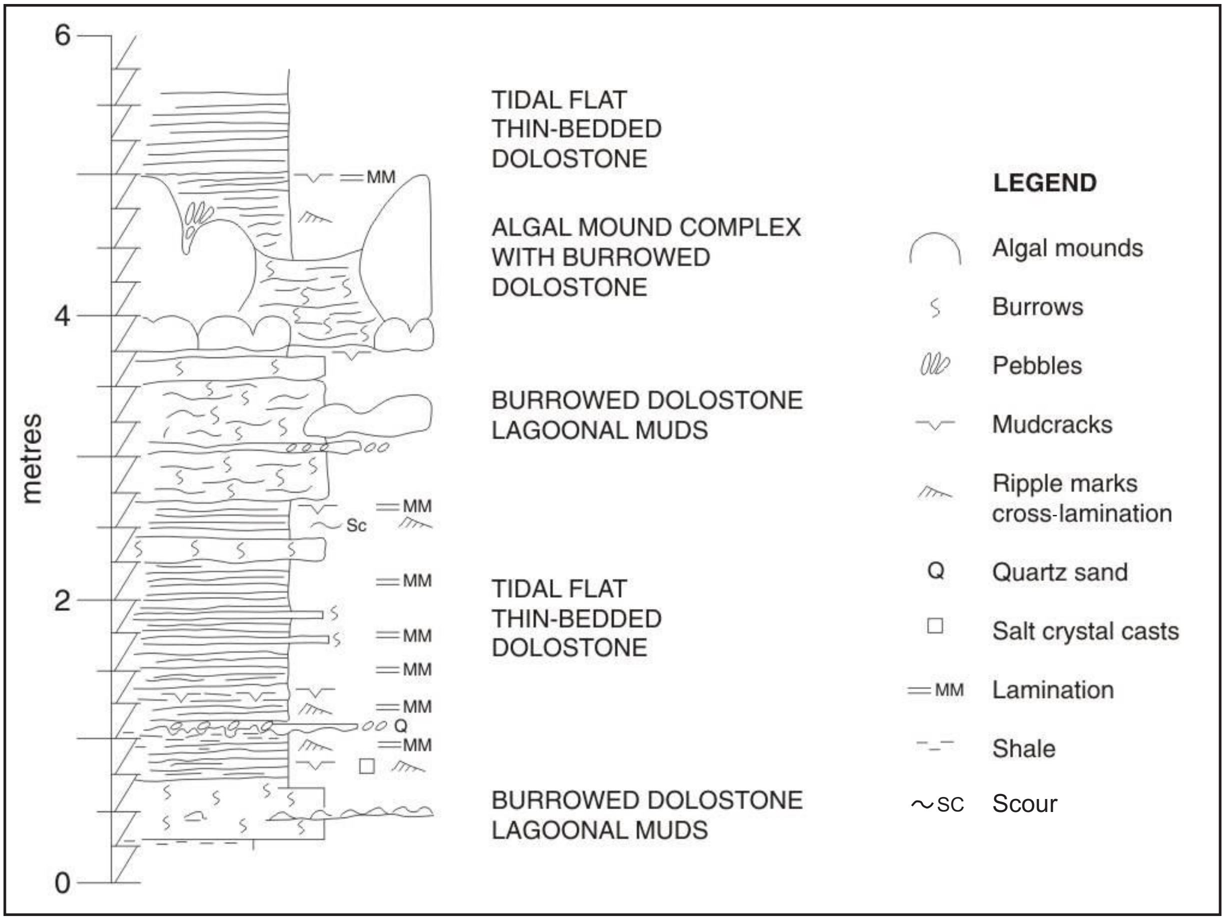

Figure 14. *A diagrammatic section of the Port au Port Group dolostone succession at Long Bottom and Meany's Point.*

crested ripple marks; the crests trend almost due east similar to ripples in the general succession. This, plus the slight oval shapes to some mounds, probably implies that gentle currents likely moved through the complex in a north or south direction.

Buff- to yellow-weathering, thin-bedded dolostone that typically breaks into flagstones due to thin shale partings (Figure 15E) are best observed south of the mounds on the west shore of Long Bottom. The facies also infill the upper part of the intermound areas overlying burrowed dolostone and draping, at first, against and later, over the mounds. Generally flat bedding planes also exhibit small undulose to bumpy features, the latter resembling small pin-cushion and kidney-shaped protuberances of likely algal origin. The dolostone also exhibits small straight ripple marks, flat lamination, crosslamination and irregular intraformational conglomerate layers (Figure 15F and G). The latter that consist of small dolomudstone rip-up pebbles locally mixed with scattered quartz sand grains rest upon irregular erosion surfaces. Angular shrinkage cracks and mudcracks cut green-grey shale partings and are locally associated with moulds of salt hopper crystals (Figure 15H), some of which have skeletal TV-aerial-like structures protruding from the crystal corners.

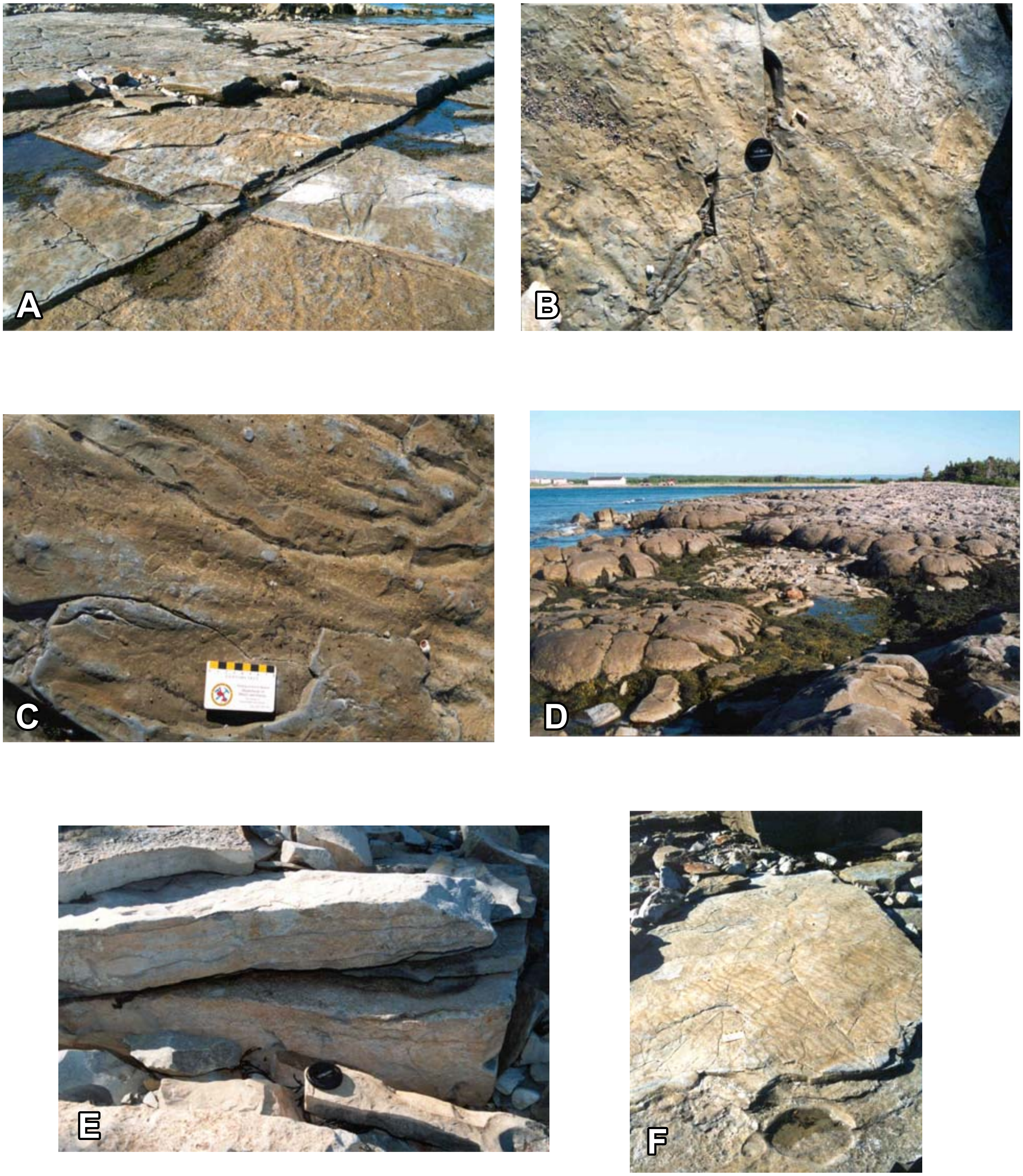

Figure 15. *Sedimentary structures of the Meany's Point–Long Bottom section. A) Flaggy, dark-grey, burrowed subtidal dolostone; B) Simple tubular burrows on a bedding plane of dark-grey, dolostone; C) Wrinkled bedding plane suggesting microbial mat (*see *Figure 15I for comparison); D) Algal mound cluster separated by irregular low areas once occupied by thin-bedded dolostone; E) Flaggy, thin-bedded, tidal-flat dolostone; F) Thin-bedded dolostone with small straight ripple marks.*

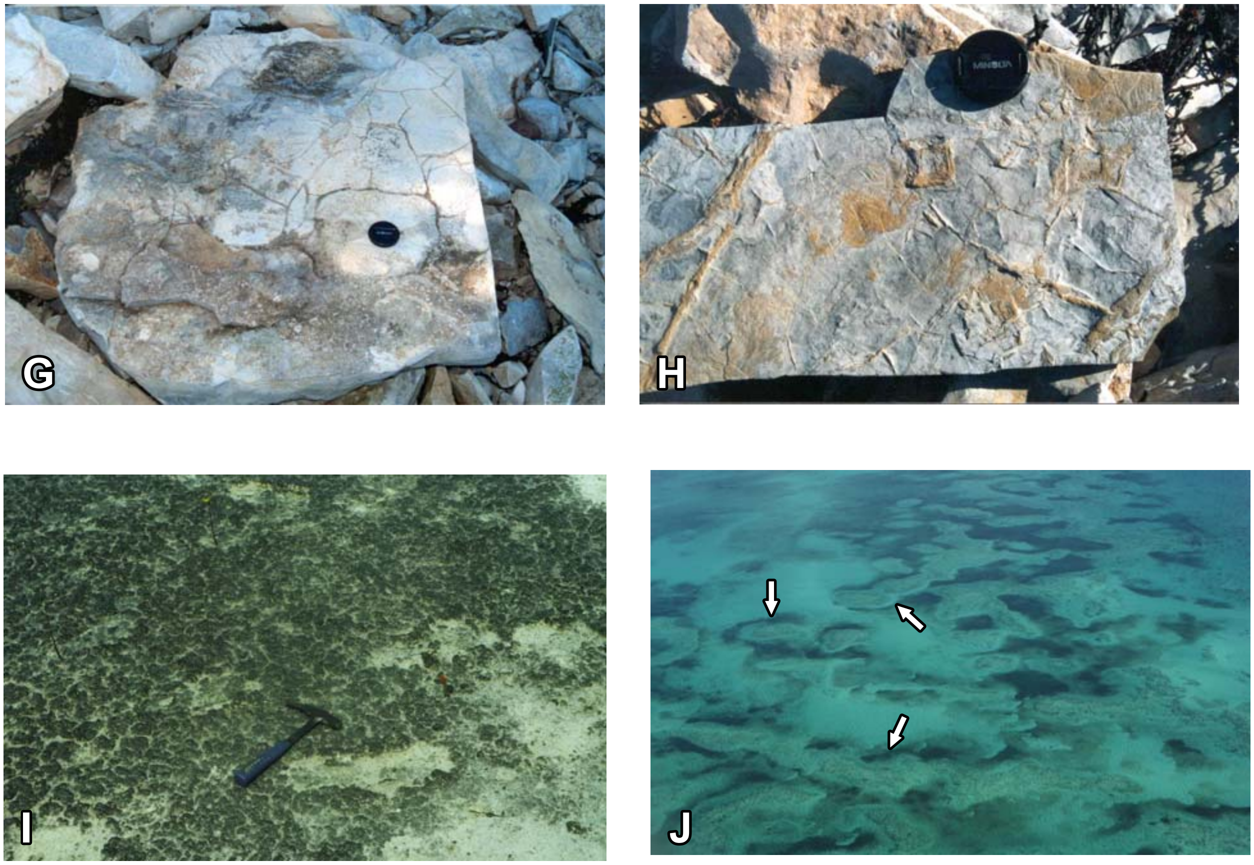

Figure 15. (continued) *G) A flagstone showing large mudcracks overlying a conglomerate of small dolostone pebbles; H) Salt hopper crystals casts (below lens cap) and shrinkage cracks on dolostone flagstone; I) Wrinkled pin-cushion algal mat, tidal flat, Bahamas. Compare Figure 15C; J) A Bahaman shallow lagoon from the air showing irregular submerged, sediment mounds (outlined by dark rims, arrows) surrounded by open mud-covered seafloor – this irregular pattern of mounds distribution may be a possible analogue for the Meany's Point mound complex. Dark areas probably marine grass. (Note: Figure 15A to D – Meany's Point; Figure 15E to H – Long Bottom; Figure 15I and J – Bahamas, 1984.)*

Depositional Environment of the Port au Port Group, Bird Cove Area

The thin stratigraphic succession at Meany's Point provides a short snapshot of the Newfoundland shelf in the Dog Peninsula area during the Late Cambrian (about 500 to 495 Ma). The shelf at this time is still considered to be a ramp (James *et al.*, 1989) but because shallow-water, Late Cambrian rocks can be traced throughout western Newfoundland, it is clear that it was much wider in extent than the Forteau ramp and perhaps even a platform by the time the Berry Head Formation was deposited. Laurentia, in the Late Cambrian and continuing into the Early Ordovician, was covered by a shallow epicontinental or epeiric sea that stretched into the deep interior of the continent leaving only a relatively small core of exposed Canadian Shield as land. High levels of atmospheric CO_2 and warm air and ocean temperatures, typical of a greenhouse earth, helped the epeiric seas to be vast natural factories that produced carbonate rocks (James and Wood, 2011). The vast, shallow-water carbonate shelf that surrounded and blanketed much of interior Laurentia is known as the **Great American Carbonate Bank**.

In the Bird Cove area, features such as the algal mounds, hopper salt crystals and the paucity of shelly fossils and bioturbation all suggest that the shallow seas were saltier than normal seawater. The seawater hypersalinity discouraged most marine animals while at the same time it provided ideal conditions for the proliferation of cyano-bacterial microbial algal mats as in the tidal flats of the Bahamas today (Figure 15I) and algal mounds of sheltered marine embayments of Shark's Bay, Western Australia. The thickness of the cycles and the limited height of the algal mounds suggest that water depths and the tidal range of the ancient shoreline was probably only a metre or two, *i.e.*, microtidal (Pratt, 2011). This, the fine-grained muddy carbonate sediment and the high salinities, imply that the area was likely isolated, and possibly far away from, the open ocean of that day.

The seascape of the time would likely be nothing like that known today. Many tens or perhaps hundreds of kilometres away from the open ocean, the shelf should be envisioned as a broad expanse of shallow seaways and interconnected lagoons peppered with bank complexes of algal mounds and low, featureless, tidal-flat islands. Microbial mats would have been common in both subtidal and intertidal flat settings similar to tidal flats in the Bahamas today (Figure 15I). The rise and fall of tides would likely be almost imperceptible. The mixture of sheltered setting, low-energy conditions, warm water, hypersalinity, dominance of algal mound builders and lack of a calcareous shelly fauna all conspired to produce only carbonate mud. The haphazard development of skeletal–algal mounds surrounded by lime-mud-filled lows seen in the shallow, sheltered, central lagoon of the Bahamas Banks may serve as an appropriate image of the Cambrian lagoon at Bird Cove (Figure 15J).

The burrowed, fine carbonate mud was probably laid down in a lagoonal or very shallow subtidal shelf setting. The simple burrow types, the microbial mats on the seafloor and the absence of fossils suggest that the salinity was probably too high to be hospitable for most marine animals. The few burrowers and perhaps salinity-tolerant inarticulate brachiopods are the exception. The thrombolitic mounds grew submerged in the shallow lagoon. The lack of carbonate sand bodies in the succession and of erosion of the mounds suggest generally quiet conditions prevailed. The flat to gently pillowed top of the larger mounds, however, implies that they may have grown into the intertidal zone to be exposed at low tide. The smooth shallow furrows that occur between each pillow and continue down the steep sides of some mounds may have been fashioned by runoff of seawater during falling water of the tidal cycle.

The thrombolitic–boundstone mounds characterized by an unstructured clotted fabric in contrast to a laminated structure as seen in stromatolitic–boundstone mounds, formed by the binding activity of blue-green cyano-bacterial algae and other microbes. The latter, not only trapped fine sediment, but also contributed to the precipitation of fine carbonate mud from the seawater. The mounds were rapidly lithified to form rigid, hard structures like modern analogues found in the tropical shorelines of the Bahamas and Shark's Bay, Western Australia. There in Western Australia, rigid limestone mounds dressed in a skin of living microbial mat thrive in shallow lagoons and in the lower part of the intertidal zone, where they are commonly exposed at low tide. In the Bahamas, the thrombolite mounds occur in tidal channels. It is also worth noting that hard, lithified thrombolite mounds and domal microbialite mounds are also found today along the shores of shallow inland water bodies in Western Australia, *e.g.*, Lake Clifton and Pink Lake. Both are cut off from the nearby ocean and affected by changeable salinity and water conditions. All together, it suggests, that microbialite mounds can thrive in a wide range of settings, salinities and energy conditions (Burne and Moore, 1987; living examples and excellent illustrations of stromatolite and thrombolite mounds in their natural settings can be found on websites (*e.g.*, YouTube, search site for stromatolites in Australia).

The larger complexes of coalesced mounds at Meany's Point were probably algal banks that were dispersed widely in irregular clusters across the lagoon likely seaward of tidal flats. The sinuous, mound-free areas that wander between the mound banks, although possibly channels, were likely just open submerged areas. The small, straight-crested ripples within the north–south trend of mounds and channels suggest that at least locally, the seafloor gradient was north–south. The small ripples in these areas suggest gentle currents (likely microtides) flowed between the maze of mounds. Only the flat disc-like pebbles locally wedged into furrows between pillowed tops of the mound banks require sufficient energy at times to erode material off the seafloor and to move and flip pebbles up onto and between the mounds. This implies storm-driven currents at times may have swept through the open channel-like areas sweeping them clean of mud or sand that accumulated there during fair weather. Nevertheless, as pebbles occur on top of the mounds and in layers in the thin-bedded dolostones that drape onto and over the top of the mounds, these pebbles may have formed by local erosion in a tidal-flat setting after the demise of the mounds (*see* below).

The generally, thin-bedded dolostone was laid down repetitively over a broad, flat depositional area, such as a tidal flat. The facies infills the upper part of the intermound areas above burrowed dolostone and drapes against and over the mounds, suggests that the tidal flats, at least locally, developed in the lee of the mound complexes and gradually prograded into the lagoon burying the mounds in the process. Diurnal, seasonal and wind- and storm-dominated tides likely transported fine sediment from the lagoon onto the tidal flat, where it was stranded by receding micro tides and trapped by the growth of sticky algal mats. Bumpy surfaces of small pin-cushion and kidney-shaped protuberances indicate the importance of algal mats growing on the tidal flat and trapping sediment. The presence of small ripple marks suggests gentle wave action in shallow water and the mudcracks indicate that the higher parts of the tidal flat were exposed to the tropical sun and desiccated repeatedly for long periods. Storms periodically swept across the tidal flat and ripped up semi-consolidated sediment to form layers of pebbles lying upon erosional surfaces. Scattered quartz sand found in the pebble layer at Long Bottom, was probably wind blown and implies a source in the continental interior as well as exposure of the tidal flat. Terrigenous mud, sorted as distinct shale partings in the dolostone, also implies a steady supply of wind-blown continental dust reached the tidal flat.

Impressions of salt hopper crystals lying on mudcracked bedding planes show that the salinity of the tidal pools on the flats was at times very high. The salt crystal (also called halite) is not preserved but the hopper crystal cast marks where a crystal settled into a mud layer and left a mould when the crystal was dissolved. Halite crystals usually have a cubic form, but the square shaped, thin raft-like hopper crystal with its step-like layers indicates that the crystals grew as a thin sheet in several stages, the crystal growing rapidly at its edges only. In some cases, only the corners of the salt crystal grew, producing a crystal that has skeletal TV-aerial-like structures protruding from the crystal corners.

The salt crystals likely seeded at the surface of a shallow pool of very salty brine where the surface tension of the brine allowed the crystal to float unaided. The essentially flat crystal grew, layer by layer, to float raft-like on the pool. Disturbance of the surface tension of the brine by wind or instability of large crystals likely caused the crystals to sink into the mud at the bottom of the pool (Demicco and Hardie, 1994). Later dissolution of the crystal left an impression preserved in the mud, the mould filled with sediment to form a cast of the halite crystal. The presence of mudcracks on the bedding plane indicates that eventually the pool dried up, so perhaps it was then that the hopper crystal settled into the saline mud to be later dissolved when less salty water flooded the pool.

EARLY ORDOVICIAN ROCKS OF THE ST. GEORGE GROUP: A MORE NORMAL MARINE SHELF

Limestone and dolostone of the Lower Ordovician St. George Group occupy a narrow belt of shoreline from Pond Cove to Dog Point, separated from the Cambrian Port au Port Group by the Plum Point fault (*see* Figures 8 and 16). Two formations occur in this fault block, the Watts Bight and the Boat Harbour formations.

Watts Bight Formation

The Watts Bight Formation, a dark-grey, crystalline dolostone, is exposed in the roadcuts and along the shoreline from Pond Cove to just south of Blue Cove. The dolostone is epitomized by large thrombolite mounds and by burrowed dolostone, both of which result in a light-grey/dark-grey, strongly mottled rock. Small cavities filled by geopetal dolostone and ghost outlines of gastropods are common. The dolostone is locally rich in dark-grey to black chert nodules that locally outline the tops of some mounds; a hint of a petroliferous odour often accompanies freshly broken rock. The formation was originally a black limestone that was pervasively replaced by dolomite.

The rocks were laid down on an open, subtidal shallow shelf during marine transgression linked to rising global sea levels about 488 Ma. The Newfoundland part of the Laurentian margin was characterized by copious growth of large algal mound complexes, the mounds dominated by the activity of sediment trapping cyanobacterial algae but unlike older Cambrian mounds, joined by primitive metazoan reef builders, such as coral, stromatoporoid and sponge; robust gastropods and cephalopods also benefitted from life on, and close to, the mounds that responded to warm clear seas, an oceanic influence and locally strong currents.

Boat Harbour Formation

Rocks of the Boat Harbour Formation underlie the western tip of the Dog Peninsula, Grave Point and Wood Hill near Brig Bay, the southeast shore of Old Ferolle Harbour, Old Ferolle Island and Moyac Island and other small islets nearby Northern Pass. In western Newfoundland, the formation consists of repetitive, metre-scale, shallowing-upward cycles of interbedded limestone and dolostone. The limestone includes burrowed and fossiliferous limestone, stromatolitic and thrombolitic mounds, fossiliferous and pebbly grainstone and laminated limestone; dolostone is generally laminated. Chert is locally common. Trilobites, brachiopods, gastropods and cephalopods are known from this unit in the area (Figures 17A, 18, 19 and 20). The cyclic sediments were deposited in shallow shelf, microbial mound complexes and tidal flats, each cycle lasting a few tens of thousands of years. To see the formation at its best exposure, the shoreline section north and west of Eddies Cove West is recommended (Knight, 1991).

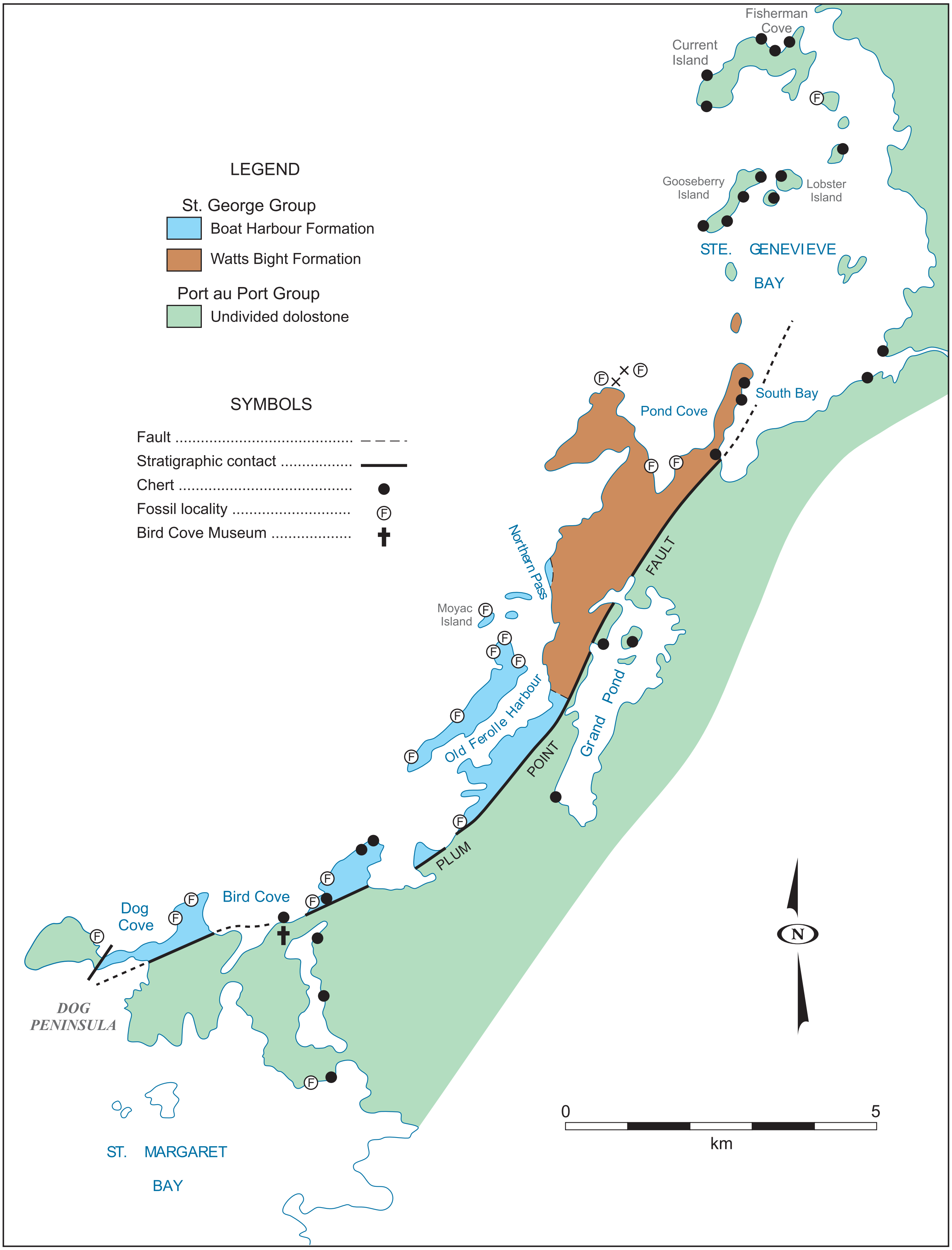

Figure 16. *Distribution of fossils and chert in the Dog Peninsula to Ste. Genevieve Bay area; based on 1978 mapping of Gary Snow (Snow and Knight, 1979).*

In the Brig Bay area, however, the rocks of this formation are generally not well exposed, and are altered by pervasive dolomitization[9]. The dolostones also host frequent bodies of breccia, some of which are mineralized by pyrite. This alteration makes this unit difficult to recognize and interpret and only a few vestiges of limestone, many of which are fossiliferous, scattered around the shoreline allow its positive identification. The pervasive dolostone is generally dull grey, massive and finely crystalline with zones of white crystalline dolomite veining locally. This style of dolomite is an indication that the succession was deeply buried and affected by warm hydrothermal fluids. The latter dissolved spaces in the buried rock into which the white dolomite crystals were later precipitated. Such dolomites are generally linked to faults along which the hydrothermal fluids migrated; the zinc mineralization at Daniels Harbour formed in such a fault-controlled, hydrothermal setting.

Limestone and Algal Mounds of the Boat Harbour Formation, Beach Point and Dog Cove

Dark-grey, thin-bedded limestone replaced by dark-grey dolostone and dolostone rich in irregular veins of white, crystalline dolomite is exposed on the west side of Beach Point. The burrowed and thin-bedded limestone is exposed as large inclined bedding planes that facilitate a walk on an ancient seafloor; dolomite replaces fossils and burrows in the limestone in this area. Well-preserved fossil snails that include the low- and high-spired gastropods *Lecanospira, Ceratopea, Euomphalus, Maclurites, Prohelicotoma and Hormotoma* can be seen locally (Figures 17A and 18). Domal stromatolite mounds – resembling those of Meany's Point – but made of dark-grey limestone, also occur (Figure 17B); most mounds in Dog Cove are replaced by dolostone. The trilobite *Hystricurus oculilunatus*, Ross 1951 (Figure 19) is recorded from the northern part of Beach Point (Boyce, 1989). Laminated and mudcracked dolostone is also common in the area. Because the succession is intermittently exposed as well as gently folded, compilation of a continuous section is impossible.

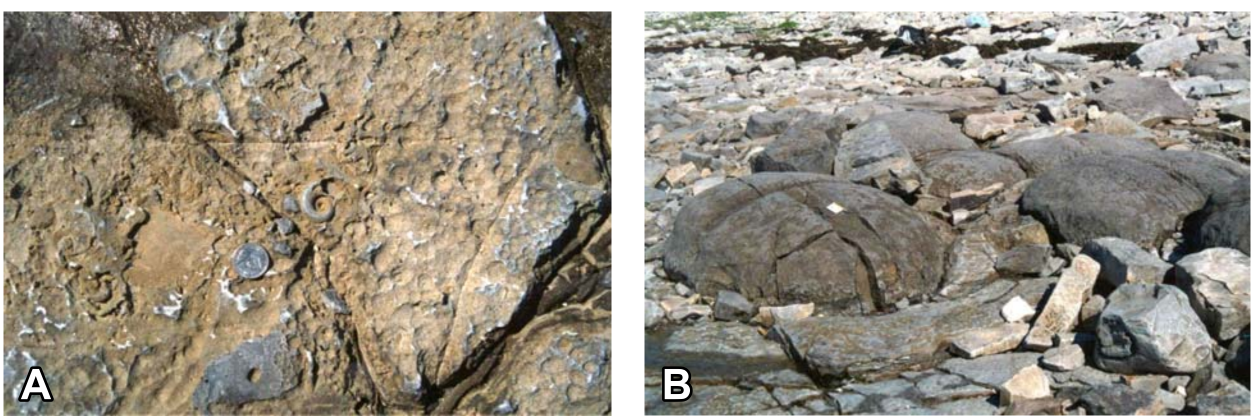

Figure 17. *Ordovician limestones of the Boat Harbour Formation, St. George Group, west side of Beach Point, Dog Peninsula. A) A low-spired gastropod (by coin) in a burrowed limestone; B) Large algal mounds; they are at least 20 million years younger than the similar Cambrian mounds at Meany's Point.*

[9] Dolomitization is a process where limestone (calcium carbonate, $CaCO_3$) is converted to dolomite ($CaMg(CO_3)_2$) during shallow to deep burial. Under favourable deep burial conditions, crystalline dolostone can become porous to form petroleum reservoirs and host mineral deposits rich in lead and/or zinc (*i.e.*, Mississippi Valley type), *e.g.*, the Daniels Harbour Mine.

Fossils of the Boat Harbour Formation, Old Ferolle Island and Adjoining Islets

Beds and remnants of limestone occur locally at a number of localities on Old Ferolle Island, Moyac Island and adjoining islets near Northern Pass (Figure 16). The dark-grey limestone ranges from burrowed and thin bedded with trilobites, brachiopods and gastropods, to pebbly limestone with trilobite fragments, to stromatolitic and thrombolitic mounds, the latter hosting trilobites and gastropods; wavy thin bedded and laminated dolomitic limestone and dolostone also occur. Boyce (1989) recorded the trilobites *Hillyardina levis* Boyce, 1989, *Hystricurus deflectus* Heller, 1954, *Hystricurus oculilunatus* Ross, 1951, *Randaynia saundersi* Boyce, 1989 and *Svalbardicurus seelyi* (Whitfield, 1889)[10] from these locations, some of which are illustrated in Figure 20; Old Ferrolle Island is the global type locality for the genus *Randaynia* Boyce, 1989. The trilobites are part of the western Newfoundland *Randaynia saundersi* trilobite zone of Boyce (1989) and indicate the host rocks belong to the middle part of the formation and lie close to the top of the Tremadocian succession in western Newfoundland.

The trilobites variably occur with the gastropod *Lecanospira* amongst others like at Beach Point (*see* Figure 18A to C), the straight cephalopod *Bassleroceras* (Figure 18E) and the single-shelled, cap-shaped gastropod (monoplacophorid) *Floripatella quebecensis* (Billings, 1865)? (Figure 18D). Articulate brachiopods, including *Diaphelasma* sp. and a very small, undetermined orthid (Figure 18F), are also part of the shelly fauna. The late Tremadocian *Floripatella* is not very common in the formation; so far, in western Newfoundland, it is only recorded from Old Ferrolle Island. It is an ancient cousin of modern day limpets and is one of the oldest known fossils of its kind; it likely grazed on algae and other microbes on the soft sea bed in the shallow subtidal and intertidal zone (Lindberg, 2009).

STRUCTURE OF THE BRIG BAY AREA: THE LEADING EDGE OF THE APPALACHIAN OROGEN

MAJOR ACADIAN FAULTS

The Mount St. Margaret and Dog Peninsula area lies at the northwestern leading frontal edge of the **Appalachian Orogen**. Three large, reverse faults, the informal Castor Pond fault, the Ten Mile Lake Fault and the informal Plum Point fault, all interpreted to have moved in the Acadian Orogen (390 to 375 Ma), occur in the area (Figures 1 and 8A). Each of these northeast-trending faults is a steeply to moderately southeast-dipping reverse fault (Figure 8B) with a northwestward sense of overthrusting, *i.e.*, older rocks moved upward to a higher structural level to be emplaced above younger rocks.

Based on geological mapping in this part of the Great Northern Peninsula (Bostock *et al.*, 1983; Knight, 1986, 1991), the vertical displacement or throw on the Ten Mile Lake Fault is calculated to be between 500 and 600 m. The Castor Pond fault that separates the uplifted Long Range Inlier from the downthrown MSM and the Highlands of St. John block probably has less throw (about 300 to 400 m). The Plum Point fault strikes northeast in a broad sweeping curve from the southwest tip of the Dog Peninsula, 17 km to Pond Cove. It separates the upper part of the

10 formerly *Paraplethopeltis seelyi* (Whitfield, 1989).

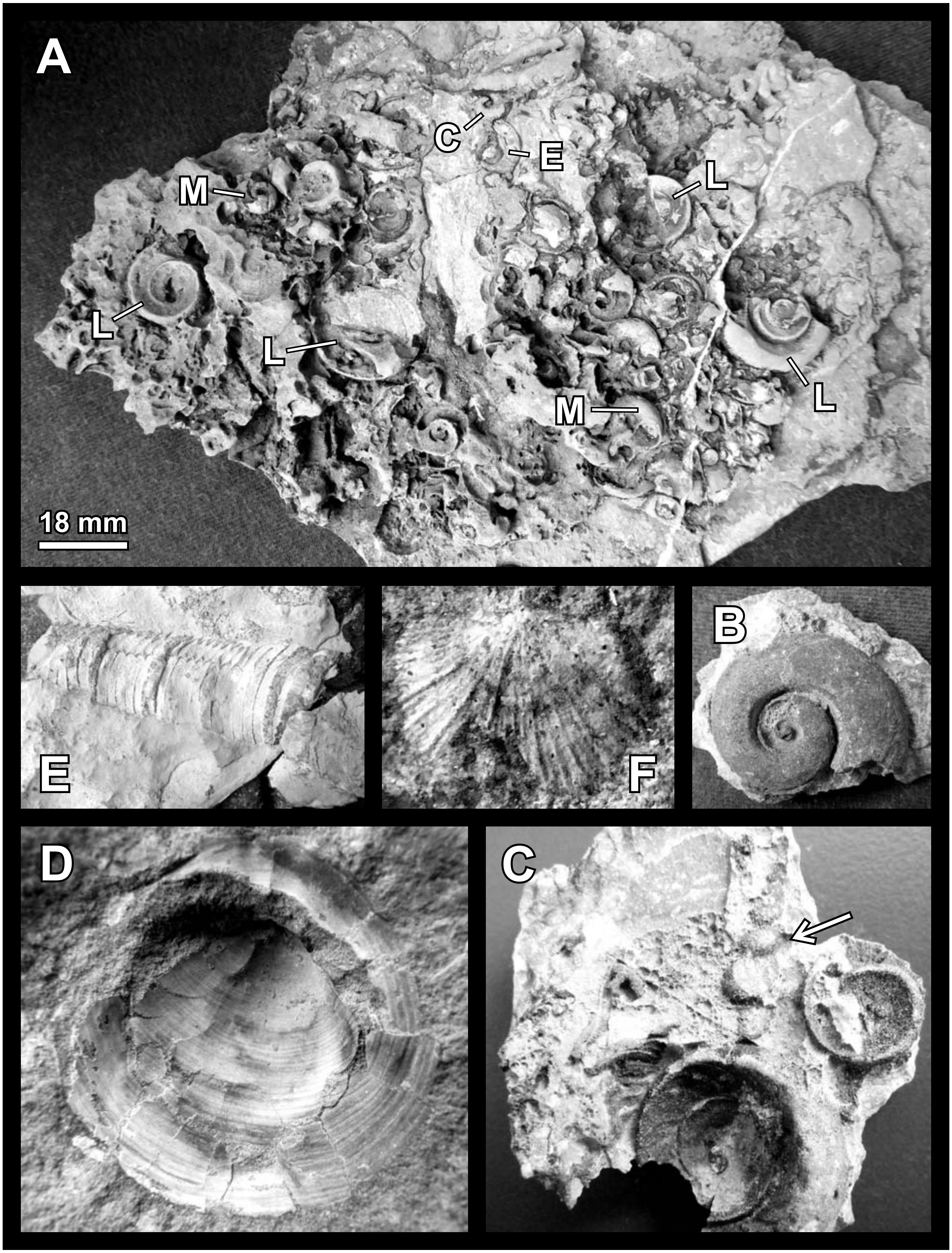

Figure 18. *Caption on opposite page.*

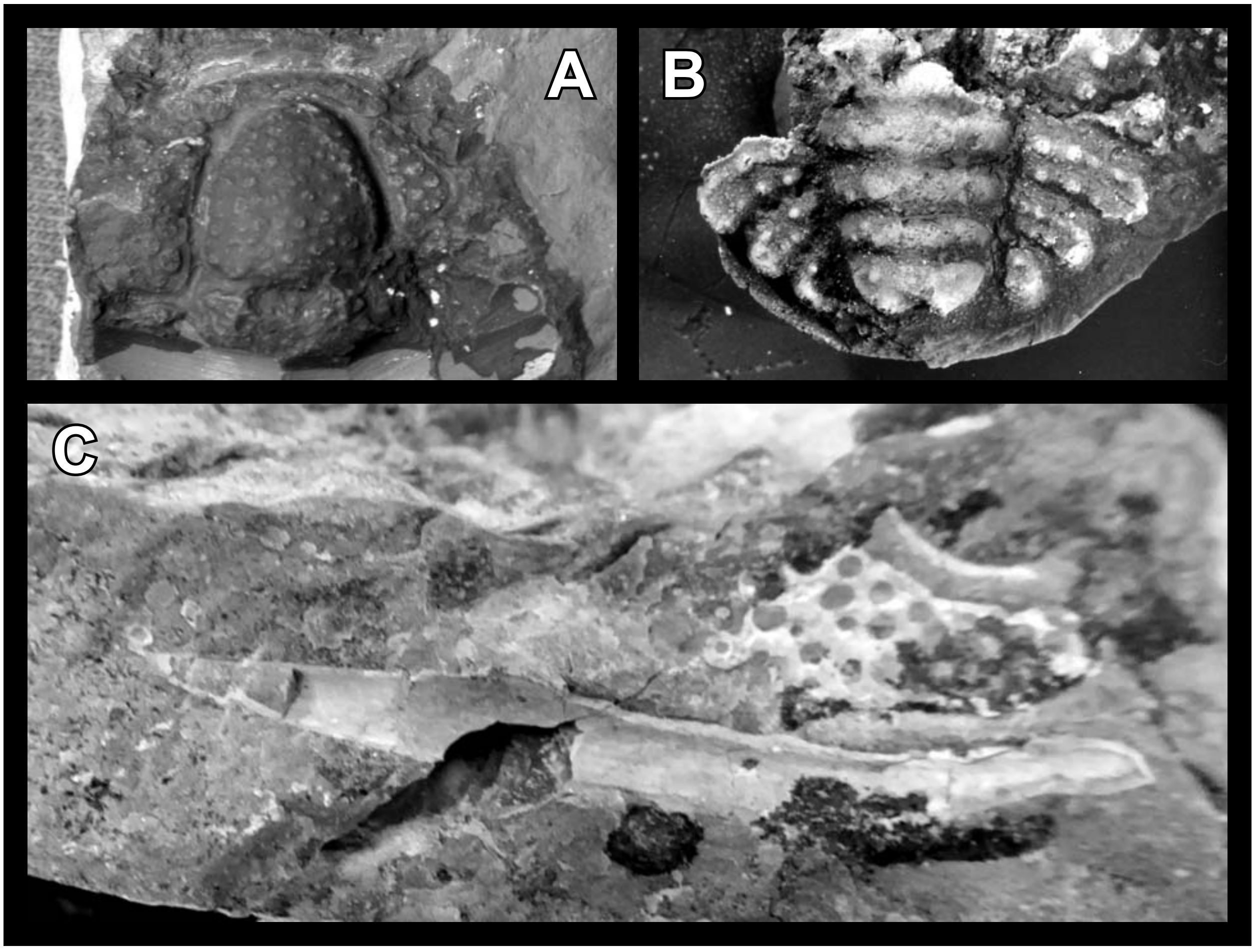

Figure 19. *Trilobites of the Boat Harbour Formation, Beach Point, Dog Peninsula:* Hystricurus oculilunatus *Ross, 1951; A) cranidium (incomplete head, about 1.5 cm wide), B) pygidium (about 1.5 cm wide), C) free cheek (about 1.5 cm long). All dimensions refer to fossils.*

Figure 18 (opposite). *Shelly fossils of the Boat Harbour Formation, Bird Cove area:* ***Molluscs: Beach Point, Dog Peninsula:*** *A) cluster of gastropods – abundant planispiral* Lecanospira *(*L*), low-spired* Ceratopea *(*C*),* Euomphalopsis? *(*E*) and* Maclurites? *(*M*) - in a largely dolomitized limestone (coin 18 mm wide); B)* Ceratopea *(4 cm diameter); C) ghost of the high-spired gastropod* Hormotoma *(15 mm long; arrow) with planispiral* Lecanospira*;* ***Old Ferrolle Island:*** *D)* Floripatella quebecensis *(Billings, 1865) (30 mm diameter), an Ordovician cousin of the modern limpet; E) longitudinal cross-section of the straight cephalopod* Bassleroceras *(4.4 cm long);* ***Articulate brachiopod****; F) pedicle valve of an unidentified orthid (3 mm wide). All dimensions refer to fossils. Fossils in Figures 18 to 20 are housed at The Rooms, Provincial Archive and Museum, St. John's.*

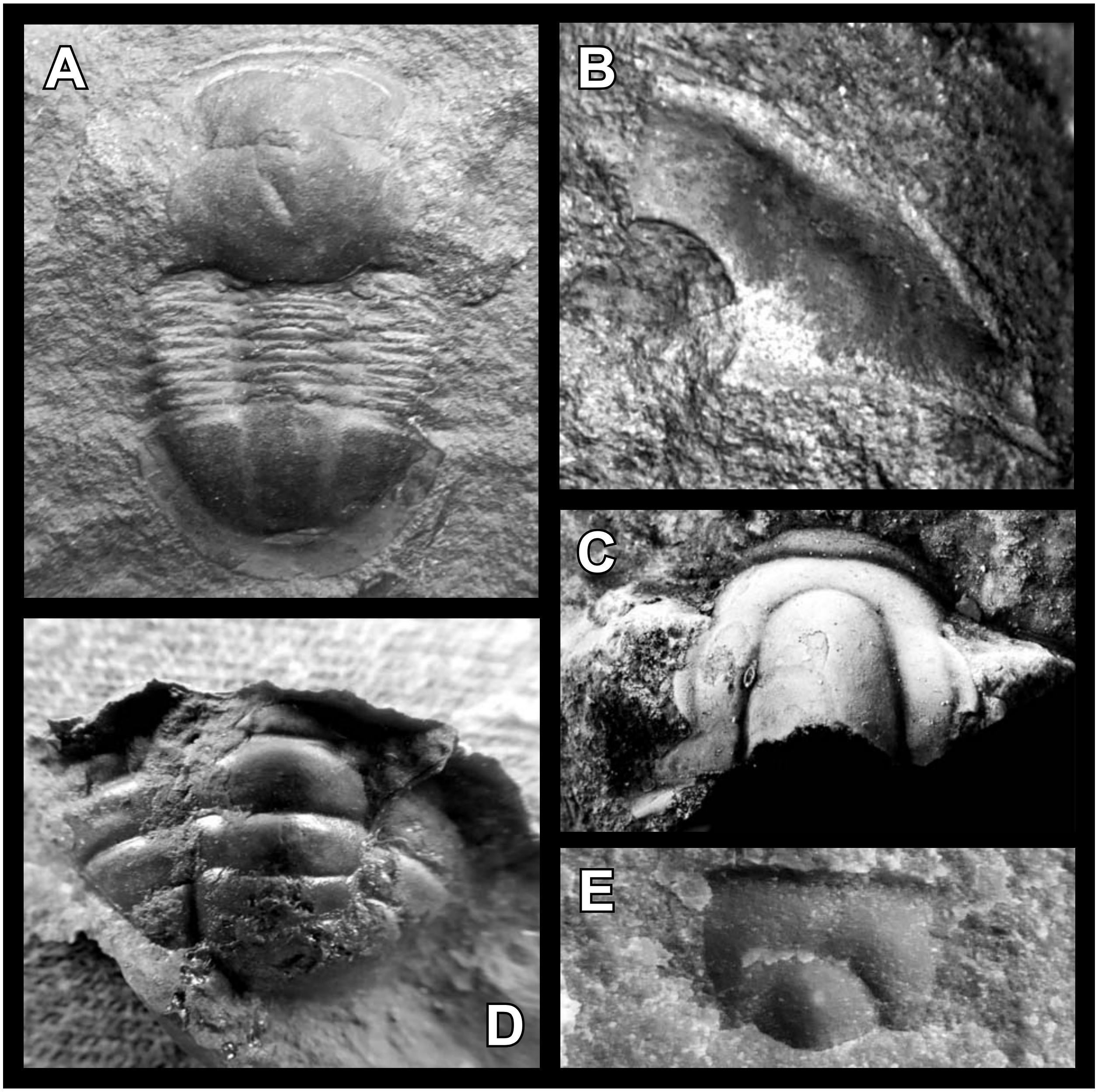

Figure 20. *Trilobites of the Boat Harbour Formation, Old Ferrolle and Moyac islands:* Randaynia saundersi *Boyce, 1989 (1.5 cm long), Old Ferrolle Island; A) type specimen; B) librigena (free cheek);* Svalbardicurus seelyi *(Whitfield, 1889), C, cranidium, D, pygidium (each about 20 mm across), Moyac Island; E)* Hillyardina levis *Boyce, 1989, cranidium (4 mm across), Old Ferrolle Island. All dimensions refer to fossils.*

Cambrian Port au Port Group to the southeast from rocks of the middle part of the St. George Group to the northwest, a throw of at least 300 m. Some small folds close to the Plum Point fault trend oblique to the main fault trend (*see* Figure 21). These folds suggest that the fault may also have moved sideways, implying wrench or strike slip movements also occurred on the fault.

Traditionally, the Appalachian structural front has been placed at the Ten Mile Lake Fault that separates uplifted Precambrian basement and Labrador Group sedimentary rocks to the southeast

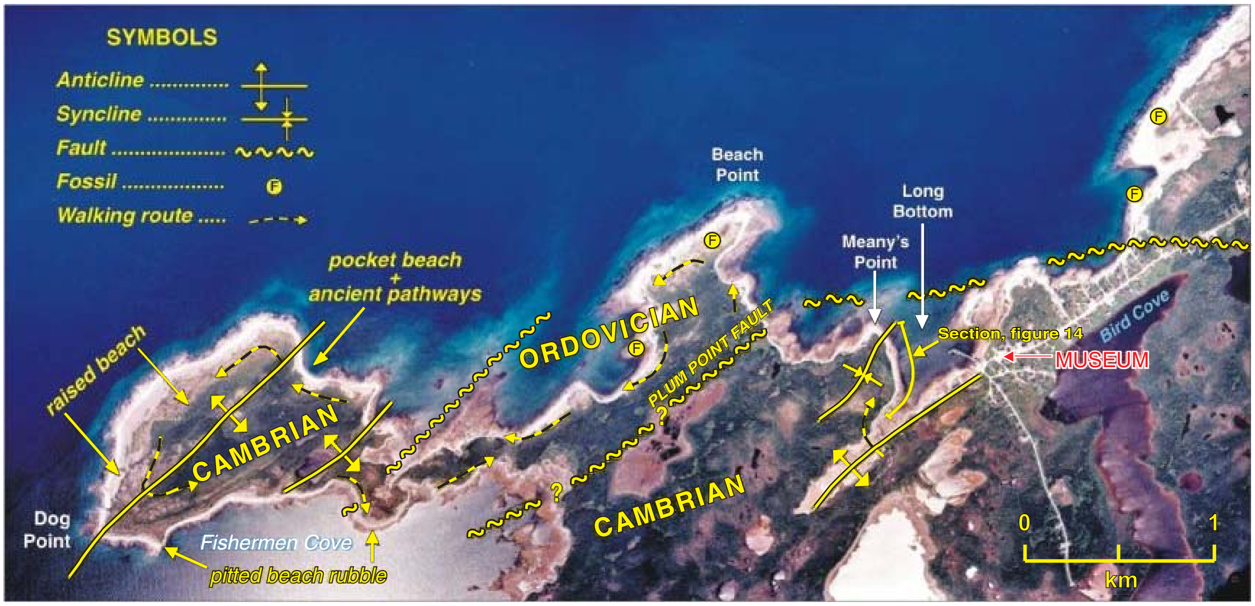

Figure 21. *An airphoto mosaic of the Bird Cove to Dog Peninsula area showing the geology and trace of the Plum Point fault. The walking route around Dog Point from Bird Cove will take between 2 and 3 hours.*

(MSM and the Highlands of St. John) from Cambrian dolostone of the Port au Port Group to the northwest (Williams, 1995). However, the common history of the three faults described here suggests that they are all Acadian faults and are, therefore, part of the Appalachian deformation. Consequently, the Appalachian front must therefore lie at the Plum Point fault.

The Plum Point Fault of the Dog Peninsula

The Plum Point fault is exposed at several shoreline outcrops along the south shore of Old Ferolle Harbour and around the Dog Peninsula (Figure 21). A narrow body of fault breccia locally marks the fault near the navigation light at Plum Point, while a deeply entrenched gash in the rocks marks the fault just north of the Bird Cove Museum. The fault is also seen at low tide on the rocky platform at the west side of Bird Cove. There, the fault lies approximately midway across the cove where it trends (strikes) northeast and dips steeply (75°E) to the southeast. Strata are folded on both sides of the fault at Bird Cove (Figure 21), and numerous northeast-trending folds deform the footwall rocks of the Boat Harbour Formation elsewhere in the area. Beds steepen close to the faults (Figure 22A) and the change of dip attitude of beds locally indicates that a number of gentle folds are present along the shoreline. Many linear fractures and minor faults (Figure 22B), mostly with the same orientation as the main fault, cut the rocks nearby in the cove.

AFTER THE ACADIAN OROGENY

By the end of the Acadian Orogeny, the Iapetus Ocean had closed as volcanic oceanic islands and distant continents collided with the Laurentian continental margin. During this long-lived process (~95 Ma), that lasted from the Late Ordovician (460 Ma) to the Devonian (390 to 375 Ma),

Figure 22. *Some structures of the Dog Peninsula. A) Steeply dipping Cambrian dolostone adjacent to the Plum Point fault in the cove south of Beach Point; B) A minor fault associated with folded Cambrian dolostone at Bird Cove. The S-shaped asymmetry of the fold and the southeast dip of the fault reflect northwestward-directed Acadian compression.*

huge masses of rocks from beneath the ancient ocean (ophiolites) had been slowly pushed into western Newfoundland, as for example the Tablelands of Gros Morne National Park and the White Hills of Hare Bay (Figure 1) near St. Anthony. Cambrian and Ordovician shelf rocks throughout much of western Newfoundland were complexly folded and faulted (*e.g.*, Canada Bay, Hare Bay, Pistolet Bay, Goose Arm) and were uplifted together with the rest of Newfoundland to form the rugged mountain belt that today forms the northeastern end of the Appalachian Mountains of eastern North America. The bays of the St. Barbe coast and the islands and coves of the Dog Peninsula and Old Ferolle Harbour lie at the very northwestern edge of the mountain front.

Although the Plum Point area provides no direct evidence for when and how often the faults moved following the Acadian Orogeny, it is probable that earthquakes shook the area during the Carboniferous (~350 to 300 Ma). Just 60 km to the east at Conche on the eastern shore of the Great Northern Peninsula folded Carboniferous sandstone and shale deposited by rivers and in lakes is confined to a fault-bounded basin. The basin was one of a series of large landlocked sediment-filled depressions in western Newfoundland and Maritime Canada that formed along the Cabot Fault system. The fault was the locus of huge wrench movements that cut through the Appalachian mountain belt (Gibling *et al.*, 2008). Non-marine sedimentary rocks fill the Deer Lake and Conche basins while Carboniferous seas invaded the most southerly wrench basin in the area of St. Georges Bay and the Codroy Valley (Knight, 1983; Hyde, 1995). At this time, the climate was tropical to savannah to desert-like as Newfoundland remained near the equator. Close, modern day analogues, both tectonically and environmentally, are provided by comparison to the valley of the

Dead Sea in the Jordon Rift Valley and Death Valley in California. Fault stresses in western Newfoundland that would have been large enough to form the rift valleys of the Conche area imply that major earthquakes certainly rocked the Dog Peninsula in the Carboniferous. For the next 300 million years, the geological history of western Newfoundland is largely obscure, although it is probable that the mountainous island was eroded under tropical, temperate and ice-age climates.

LOCAL GEOLOGY AND ARCHAEOLOGY

LOCAL SOURCE MATERIAL USED BY ANCIENT SETTLERS

Ancient peoples who inhabited the Bird Cove area, millennia before Europeans settled in the area fashioned their tools from rocks for every day use (Hartery, 2001). Several stone tools have been excavated from Cains Garden and other local sites. The rock materials used include chert, slate and grey sandstone. Chert, which dominates recovered material, is locally common but materials such as the slate or the grey sandstone is not endemic to the area and must have come from other parts of western Newfoundland or farther a field. Much of the chert used in local artefacts is not endemic to the area, however. Radiolarian chert, a chert with round radiolarian microspheres, from the Cow Head Group, western Newfoundland, is the primary lithic material used by recent Indian cultures (Cow Head Complex *ca.* 1800 to 1100 yrs BP) in western Newfoundland and Québec (Hartery, 2001). Similar radiolarian chert is unknown in the Bird Cove area, although local chert has yet to be studied systematically and microscopically.

The following discussion summarizes the distribution of chert in the Bird Cove area and adjoining bays within 10 to 15 km of the Dog Peninsula. Since the earliest settlers likely roamed along the shoreline, it is assumed that they would have been familiar with the nearby shore as well as islands of Old Ferolle Harbour and Ste. Genevieve Bay. As the first explorers of the seashore and ponds of the area, they had the advantage of abundant chert weathered from host rocks over the many thousands of years, as well as material transported and abandoned by glaciers. Harvesting this material only required the ancient inhabitants to pick it off the beaches and rocky outcrops.

CHERT IN THE BIRD COVE AREA

Chert nodules, irregular masses and thin sheets of white, grey and black, cryptocrystalline chert, locally speckled with dolomite microcrystals, occur extensively in the Bird Cove area. They occur close to the seashore not far from local archeological sites as well as on the sides of ponds such as Grand Pond and farther afield on the many islands that occur in Ste. Genevieve Bay and in the vicinity of Old Ferolle Island. The chert is found in dolostone of the Berry Head Formation, upper Port au Port Group, and the Watts Bight and Boat Harbour formations of the St. George Group (Figure 16).

In the immediate vicinity of Bird Cove, collectable chert occurs in the Boat Harbour Formation on the shore just north of the Bird Cove community (Figure 21), at Grave Point and in Brig Bay as well as on nearby ridges and at Woody Hill. Near Bird Cove, the chert is associated with dolomitized stromatolitic mounds. It is more abundant on the shore and on small ridges on the west side of Woody Hill where it occurs in thin-bedded and laminated dolostone. At Grave Point, large black chert nodules up to 30 cm in diameter occur in laminated dolostone. Around the shores of Brig Bay, black chert is a significant part of fault breccias and other breccia bodies and is also associated with stromatolitic dolostone mounds. As these chert-bearing Ordovician rocks lie just north of and in the footwall of the Plum Point fault, much of the chert is fractured, causing large masses of chert to disintegrate into small fragments.

Light-grey, fine-grained dolostone of the Cambrian Port au Port Group (in the hangingwall of the Plum Point fault) also hosts chert. Only the uppermost formation in the group, the Berry Head Formation, is known to host chert. The chert-bearing, light-grey dolostone occurs at a point and on adjoining islands at the north end of Grand Pond (Figure 16) and in light-grey, locally stromatolitic, dolostone on the east shore of the narrow pond just south of Bird Cove.

Dolostone rich in chert nodules belonging to the same Cambrian formation lines the shore of many islands in Ste. Genevieve Bay (Figure 16). Chert is abundant at the southwest end and around Fishermans Cove on Current Island, and on Lobster and Gooseberry islands. There, the chert is associated with large stromatolitic to thrombolitic mounds composed of dark-grey, crystalline dolostone interbedded with the light-grey dolostone. The chert ranges from white, grey to black. Chert is also abundant on the points either side of the mouth of Ste. Genevieve River where it occurs within burrowed and stromatolitic dolostone.

The best examples of chert in the Watts Bight Formation occur on the west shore of South Bay in the south of Ste. Genevieve Bay. Dark grey and white, it occurs in stromatolite mounds as well as in bedded zones, 30 to 50 cm thick. A brecciated chert bed, 1 to 2 m thick overlying mounds also occurs in the same vicinity.

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