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GEOLOGICAL ASSOCIATION OF CANADA
NEWFOUNDLAND SECTION

"BURIN PENINSULA '88"

FIELD GUIDE AND ROAD LOG

for the

OCTOBER 14-16, 1988

FALL MEETING

LEADERS: R.N. HISCOTT, C.F. O'DRISCOLL, C.J. COLLINS,
S.J. O'BRIEN AND MINWORTH STAFF

BURIN PENINSULA'88 FIELD GUIDE

Preamble

The Burin Peninsula has long been regarded as a critical area in the geological framework of Newfoundland, and is known for both its mineral deposits, and for its superb exposures of Late Precambrian and Cambrian Avalon Zone rocks. Our trip will embrace a variety of themes ranging from fluorite and gold mineralization to sedimentology and ichnology. All, however, are unified by the fact they are topical issues of considerable economic and scientific importance. On Day one, for example, we will visit the St. Lawrence Fluorspar Mine, an operation which has only recently reopened and which will play a vital role in the economic development of the Burin region. On Day 2 we will examine the Precambrian-Cambrian boundary section on the west side of the Peninsula. This section is the current focus of world-wide scientific interest and is one of three candidates under consideration as the International Boundary Stratotype. Finally we will visit the gold mineralization of the Hickeys Pond belt; an area that exemplifies the increasing importance of gold exploration in the Province and the potential of the Burin Peninsula for further economic development.

R.J. Wardle
President
GAC Nfld. Section

ITINERARY

Day 1 (Saturday Morning): St. Lawrence fluorite deposit - tour of surface, underground operation and mill. Stops 1-1 to 1-5 (Figure 1).

Day 1 (Saturday afternoon): Examination of the Precambrian-Cambrian boundary stratotype. Stops 1-B, 2-A, 2-B (Figure 1).

Day 2 Granites, volcanic rocks and alteration zones of the northern Burin peninsula. Stops 2-1 to 2-5 (Figure 1), Stop 2-6 (Figure 27).

CARBONIFEROUS

[10] 10a, *granite;* 10b, *red sediments*

DEVONIAN

[9] *Basaltic flows and felsic pyroclastics*

[8] 8a, *mainly granite;* 8b, *mainly diorite and gabbro*

[7] *Purple, red, brown and gray sandstone, siltstone, shale, conglomerate, and limestone*

CAMBRIAN

[6] *Undivided lower, middle and upper Cambrian rocks.* 6a, *quartzite;* 6b, *red, gray, green and black shale; red, gray and pink limestone*

PRECAMBRIAN

[5] 5a, *mainly granite and granodiorite;* 5b, *mainly diorite and gabbro*

[4] *Gray sandstone, shale, siltstone, conglomerate; minor red sandstone, white quartzite (includes* MUSGRAVETOWN GROUP *and equivalents.*

[3] 3a, *felsic and mafic volcanics;* 3b, *tuffaceous sediments;* 3c, *rhyolite porphyry;* 3d, *red sandstone and conglomerate (includes* MARYSTOWN GROUP, *parts of* CONNAIGRE BAY *and* LONG HARBOUR GROUPS *and the* BULL ARM FORMATION *of the* MUSGRAVETOWN GROUP)

[2] *gray and green sandstone, shale, conglomerate* (CONNECTING POINT GROUP)

[1] *basaltic pillow lava, pyroclastics and sediments; gabbro* (BURIN GROUP)

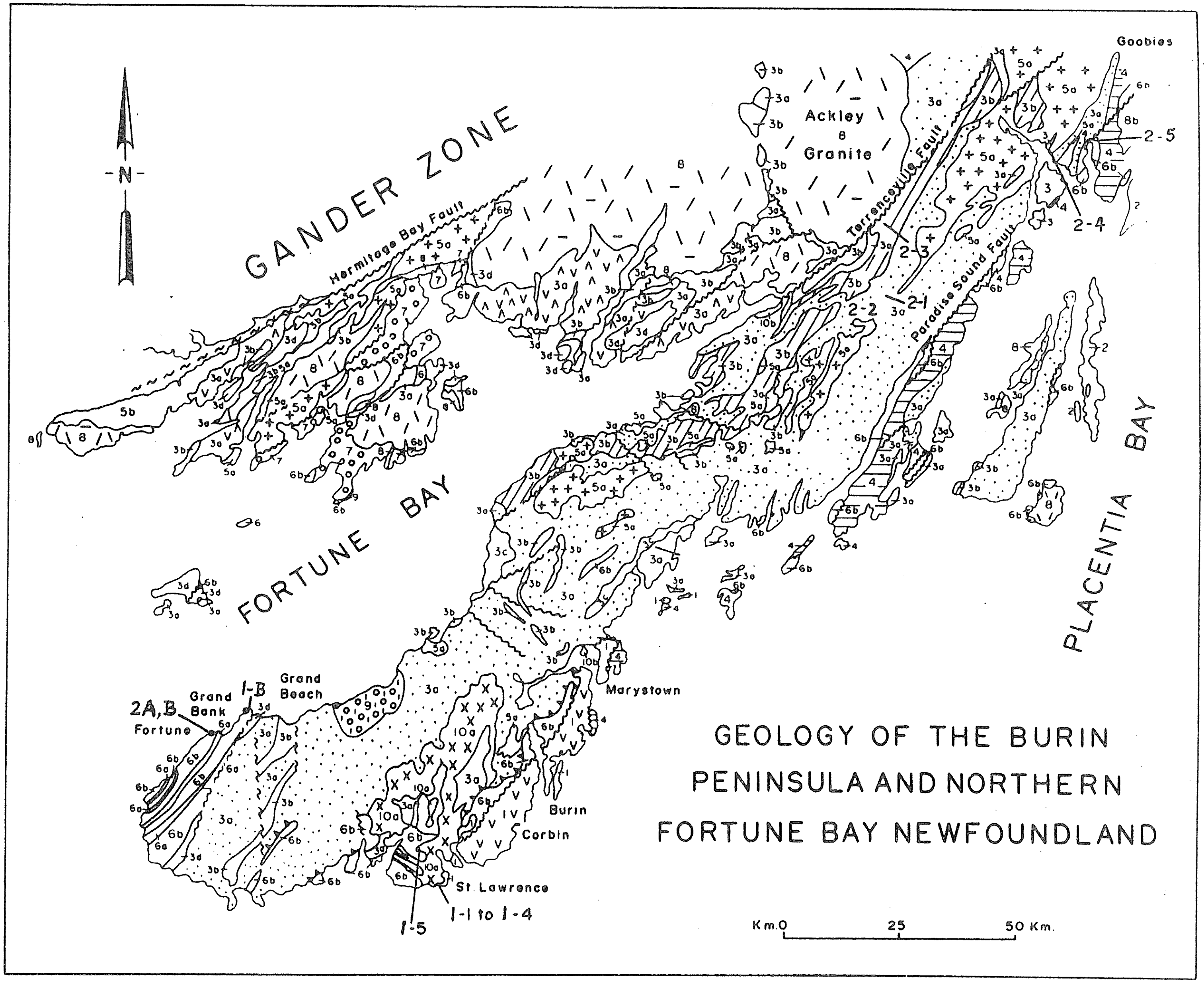


Figure 1 . Geology of the Burin Peninsula (O'Brien and King, 1982).

AVALON ZONE: AN OVERVIEW

The Avalon Zone in the Newfoundland type area comprises the late Precambrian (760 Ma to 570 Ma) volcanic, plutonic and sedimentary rocks and overlying Cambro-Ordovician strata that lie to the southeast of the Dover and Hermitage Bay faults (Figure 2). Its oldest rocks comprise a late Riphean mafic volcanic - volcaniclastic succession associated with oceanic gabbro and carbonate olistostrome (Strong et al., 1978). After 620 Ma, the zone was the site of widespread, dominantly bimodal volcanism and comagmatic plutonism, products of which are hallmarks of the zone throughout the orogen. Flyschoid sedimentation accompanied and followed volcanism in the late Vendian (Williams and King, 1979; O'Brien, 1987; Knight and O'Brien, 1988). Shoaling of these marine basins gave way to deltaic and ultimately alluvial clastic sedimentation in latest Vendian time (e.g. Smith and Hiscott, 1984, King, 1980, in press). A widespread erosional unconformity separates the Precambrian succession from overlying Tommotian to Arenig platformal sedimentary rocks, containing distinctive Acado - Baltic trilobite fauna (e.g. Hutchinson, 1962).

NOTE: A detailed discussion of the latest Precambrian and Cambrian succession is given in that part of the guide dealing with the Precambrian-Cambrian boundary.

The Avalon zone in Newfoundland records at least 200 Ma of crustal evolution prior to the Cambrian. Although the oldest dated rocks from the zone have yielded U/Pb zircon ages of 760 +/-2 Ma (Krogh et al., 1983, in press), the main Avalonian igneous pulse represents a period of only about 50 Ma in the latest Precambrian. The subsequent marine and terrestrial sedimentation continued from at least 600 Ma to the end of the Precambrian. In the lower Paleozoic, sedimentation continued for almost 100 Ma, from the Tommotian to the Arenig; there is no record in the on-land Avalonian stratigraphic column of post-Arenig - pre-Devonian rocks. By the Devonian, and through the Carboniferous, facies similar to those forming elsewhere in the orogen were deposited within the Avalon Zone. Thus, whilst the Avalon Zone in Newfoundland records 400 Ma of geological history, its characteristic lithofacies formed within a relatively short time span (Figure 3).

Extent:

On land, the Avalon Zone extends eastward from the Dover and Hermitage Bay faults for approximately 200 km to Cape Spear. Offshore geological and geophysical data (e.g. Haworth and LeFort, 1979) show that the zone continues eastward to the edge of the continental shelf. The total width of the zone is more than twice the width of the remainder of the orogen. Late Precambrian granitoid rocks have been collected from the Virgin Shoals and from the Flemish Cap, approximately 700 km southeast of the Avalon Peninsula (Lilly, 1966; Poole, 1967). Offshore magnetic and gravity data (Haworth and Lefort, 1979; Zietz et al., 1980) indicate the zone extends south of the Collector Anomaly, a major east-west-trending magnetic high located approximately 200 km south of the Avalon Peninsula. The onland trace of this feature in Maritime Canada corresponds to the Chedabucto Fault, the boundary of the Avalon and Meguma zones. F

Magmatic rocks:

The widespread occurrence of foliated, hornblende and biotite-bearing, calc-alkaline granite and granodiorite plutons is one of the most distinctive

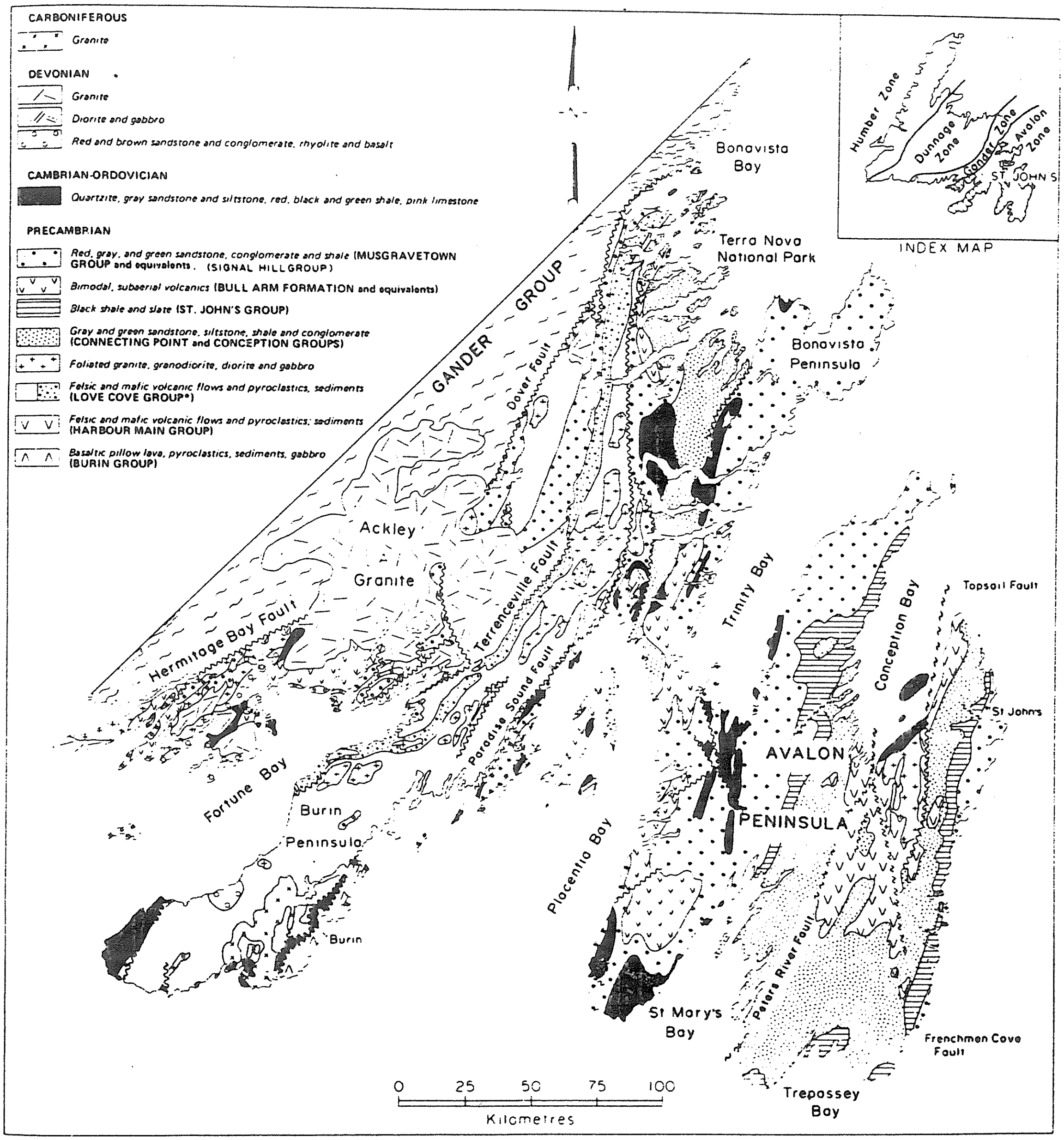


Figure 2. Simplified geological map of the Avalon Zone in Newfoundland (after O'Brien and King, 1982).

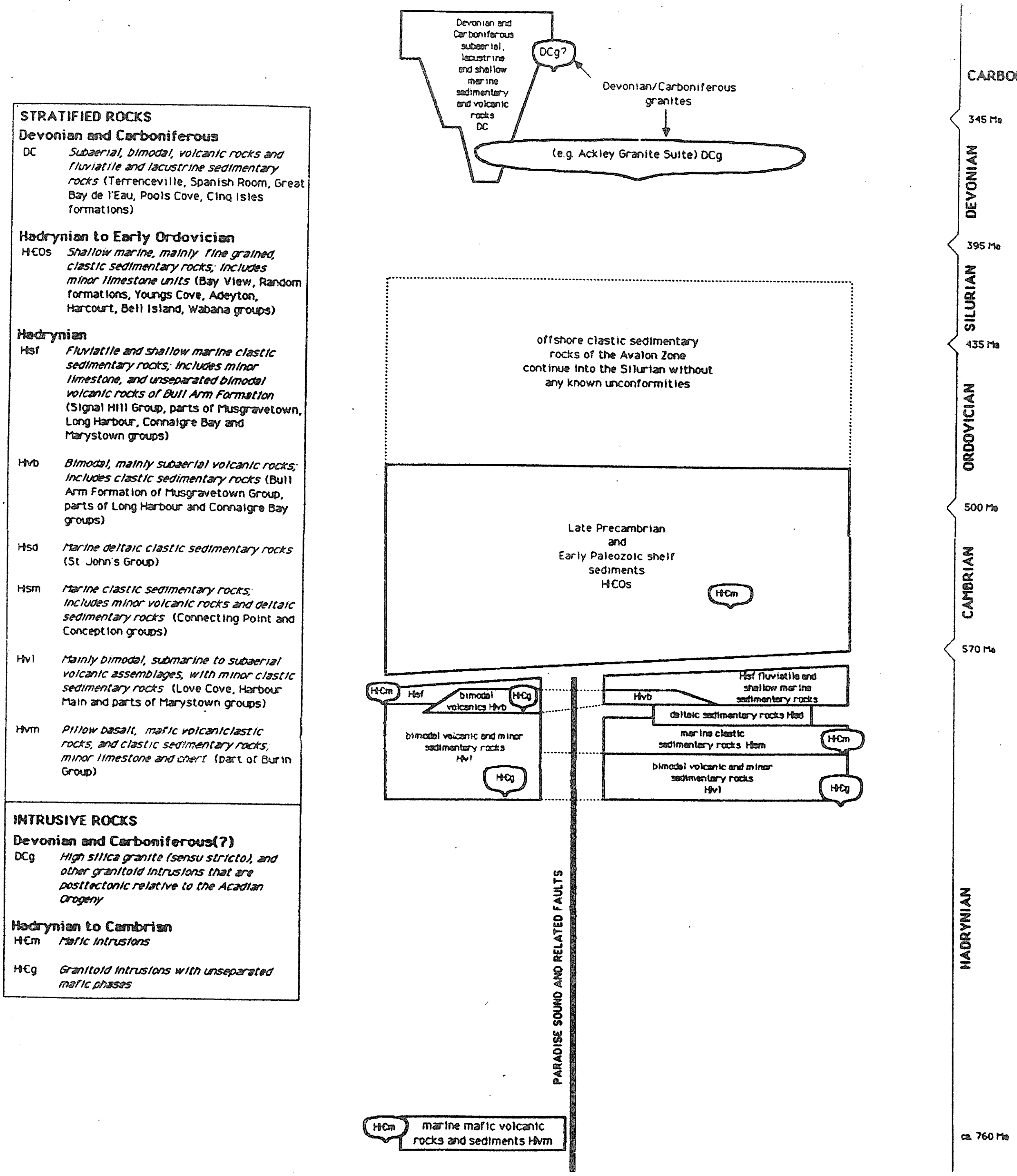


Figure 3 : Stratigraphic nomenclature, lithology and relative ages of major geologic units in the Avalon Zone in Newfoundland.

aspects of the Avalon Zone in Newfoundland, and of its correlatives in North America and elsewhere. In Newfoundland, these rocks form the core to each of the main Avalonian volcanic massifs, and are considered to be comagmatic with these volcanic rocks (e.g. Dallmeyer et al., 1981a, Figure 2). The granitoid rocks are unconformably overlain by fossiliferous Cambrian strata on the Avalon Peninsula, where they have yielded a precise U/Pb isotopic age of $620^{\pm 2}$ Ma. (Krogh et al., in press). On the Burin Peninsula, the granitoids form a number of elongate, northeast-trending plutons. One of these, the Swift Current Granite, has been dated (U/Pb zircon) at 590 +/- 30 Ma (Dallmeyer et al., 1981a). Plutons of similar composition intrude volcanic rocks of the westernmost Avalon Zone near and adjacent to the Dover and Hermitage Bay faults. Features common to many of these Precambrian plutons are the presence of locally extensive dioritic and in places gabbroic subphases, and evidence for coexisting mafic and felsic magmas.

In the western Avalon Zone, the stratified Precambrian rocks are also intruded by small plutons that comprise gabbro, alkali granite, peralkaline granite and related hybrid rocks (O'Brien et al., 1984; O'Brien, 1987). These plutons are spatially associated with pantellerites and commendites of late Precambrian age and have yielded preliminary U/Pb zircon ages and Rb/Sr mineral and whole rock ages of between 540 and 570 Ma (G. Dunning, J. Tuach, pers. comm., 1987).

Alkali gabbro dykes and sills intrude fossiliferous Middle Cambrian shales in the central portions of the Avalon Zone (e.g. McCartney, 1967). Shales of similar age also contain interstratified alkali basalt flows. Some of the peralkaline plutons described above may be in part related to this Cambrian magmatism.

Mid-Paleozoic granitoids of the Avalon Zone are typically metaluminous, A-type biotite granites that host little or no tectonic fabric. These rocks locally intrude fossiliferous Devonian strata. These posttectonic biotite granites are found only in the western part of the Avalon Zone; the largest plutons occur within a 50 km-wide corridor bounded in the west by the Dover Fault. These granites are common to both the Avalon and Gander Zones, and one of these, the Ackley Granite Suite, which has been dated at 360 +/- 5 Ma (Dallmeyer et al., 1983) and 367 Ma (Tuach, 1987), straddles the boundary of the two zones. The youngest known Paleozoic intrusion in the Avalon Zone is a high-level, hypersolvus riebeckite - aegerine alaskite that has yielded a Rb/Sr isochron age of 328+/-5 Ma (Bell and Blenkinsop, 1977).

The youngest magmatic rocks within the Avalon Zone are diabase, which outcrop intermittently along a narrow, northeast-trending rectilinear magnetic anomaly that transects the entire southeastern Avalon Peninsula, and is traceable into the offshore. The dyke has been dated at 201 +/- 3 Ma (Hodych and Hayatsu, 1980) and has been correlated with other extensive Mesozoic intrusions in the Maritimes (e.g. Papezik and Barr, 1981); all are related to the opening of the Atlantic.

Tectonothermal history:

The age of the main deformation of the Avalonian rocks remains equivocal. There is clear evidence for at least two distinct episodes of regional deformation and low grade metamorphism, one of Precambrian age, the second Paleozoic.

There is general agreement that the post-620 Ma rocks were first deformed during a late Precambrian orogenic episode known as the Avalonian orogeny (Lilly, 1966, Rodgers, 1967, 1972, Hughes, 1970). At least two ages of exposed Avalonian unconformities reflect two phases of orogenesis. The marked hiatus between the late Riphean (ca. 760 Ma) oceanic plutonic rocks and younger Vendian (post-620 Ma.) successions implies the existence of yet an earlier orogenic phase that may or may not be related to the classic Avalonian.

The older of the Avalonian angular unconformities separates redbed successions from underlying marine clastic sedimentary rocks (Hayes, 1948; Anderson et al., 1950). In some areas, the uppermost part of the Vendian alluvial clastic succession is unconformable on the base of the older flyschoid clastics, and the intervening rocks, including the entire deltaic sequence, are missing (e.g. Anderson et al., 1975). Elsewhere, particularly in the western Avalon Zone, the erosional nature of the unconformity appears to be less profound, although its angular character remains clear. The younger phase of the orogeny is marked by the unconformity at the base of the Cambrian - Ordovician platformal succession. The granites and the volcanic rocks they intrude, as well as the overlying marine sediments are unconformably overlapped by these Cambrian strata.

In most parts of southeast Newfoundland, the Avalonian orogeny has historically considered to be manifest mainly by granitoid emplacement, block faulting and gentle folding, without widespread development of penetrative tectonic fabrics (e.g. McCartney, 1969; Williams et al., 1974; Dallmeyer et al., 1983). Deformation was accompanied by low grade (prehnite - pumpellyite to chlorite) regional metamorphism. Whilst it remains clear that the Avalonian event was relatively mild, recent work in the eastern Avalon Peninsula (King, 1986) and in the northwest Avalon Zone (O'Brien, 1987: O'Brien and Knight, 1988) has shown that Precambrian faulting and associated fabric development may be more widespread than originally thought.

Unconformities in the southwest part of the Avalon Zone indicate that the area was subsequently affected by widespread deformation and low grade regional metamorphism between the upper Cambrian and upper Devonian. Ar-Ar release spectra from whole rock phyllites and from mineral separates of crosscutting plutons in the western half of the zone indicate that this deformation was related to Acadian rather than Taconic orogenesis (Dallmeyer et al., 1983). In general, the Acadian deformation, which was heterogeneous in character, resulted in widespread thrusting and strike-slip faulting and upright to east-vergent folding in the western Avalon Zone, together with the development of regionally penetrative fabrics. The main movement along the zone-bounding Dover Fault is considered to be Acadian (Dallmeyer et al., 1981b). In the eastern Avalon Zone, the major effects of Acadian orogenesis are open, periclinal folds and vertical faults.

There is little evidence of orogenic activity after the end of the Acadian. Devonian stratified rocks are either undeformed, gently tilted or affected by open warps. The Devonian-Carboniferous plutonic rocks are massive and crosscut Acadian fabrics in both Precambrian and Cambrian rocks.

Fauna:

The late Precambrian marine sedimentary rocks of the Avalon Zone contain superbly preserved examples of the oldest known metazoan fossils (Anderson and

Misra, 1968; Anderson and Morris, 1982). A variety of disc- and spindle-shaped forms have been identified, including *Charniodiscus concentricus* with *Charnia masoni* (Ford, 1958). The Cambrian to Tremadocian rocks of southeast Newfoundland are faunally and lithologically the same as similar-aged rocks in Maritime Canada, the British Midlands, Wales, Scandinavia and the western Baltic, which collectively have been named the Acado-Baltic province (Henningsmoen, 1969). Its distinctive Cambrian to Tremadocian rock types are variegated shale and mudstone, black shale and subordinate limestone, which yield *Protolenus* (Early Cambrian), paradoxidid (Middle Cambrian), olenid (Late Cambrian), ceratopygid (Tremadocian) and asaphid (Arenigian) fauna. This faunal assemblage is perhaps the singularly most diagnostic feature of the Avalon Zone and global equivalents.

Geophysical signature:

The Avalon Zone has a remarkable aeromagnetic signature, defined by arcuate, sub-parallel highs and lows, which generally correspond on-land to exposures of volcano-plutonic and sedimentary rocks, respectively. The anomalies, which are between 50 to 100 km wide, vary between +400 gammas and -400 gammas, and are interpreted to reflect the presence, throughout the offshore Avalon, of volcanic ridges intruded by granitoids and separated by intervening sedimentary basins (e.g. Haworth and LeFort, 1979). Refraction data from the Avalon indicate that it is underlain by a crust of transitional nature, characterized by seismic velocities between 6.0 and 6.6 km/s (Sheridan and Drake, 1968). Recent marine deep seismic experiments off the northeast coast of Newfoundland (Keen et al., 1986) and in the Gulf of St. Lawrence (Marillier et al., in press) have indicated that, within the limits of the survey area, the Avalon Zone is a unique crustal block, characterized by east-dipping, lower crustal reflectors and short, synformal mid-crustal reflectors.

EVOLUTION OF THE AVALON ZONE

The Avalon Zone of Newfoundland and the adjacent eastern Canadian continental margin comprises the largest single segment of typical late Precambrian Avalonian crust exposed in the Appalachian - Caledonian orogen. While this region is generally considered as the type area for Avalonian geology, it contains only the later Proterozoic elements of the zone. The Helikian basement complexes and mid-late Proterozoic platformal cover successions that comprise an integral part of the Avalonian crustal segment of Maritime Canada (e.g. O'Brien et al., 1983; Figure 4) are missing in southeastern Newfoundland. Similar rocks, however, may form the unexposed basement to the later Precambrian elements that comprise the Avalon Zone in Newfoundland. The following section summarizes the salient features, only of the Avalonian type area, which were presented in the preceeding analysis, pointing out their ramifications for tectonic modelling.

Most geological and geophysical evidence in southeast Newfoundland corroborates the hypothesis that the final accretion of the Avalon Zone to the remainder of the orogen was controlled by transcurrent movements in Siluro-Devonian times. Deep seismic data support earlier notions (O'Brien et al., 1983), based in part on timing of magmatism and regional correlation, that the zone represents a separate crustal block, whose final juxtaposition with inboard terranes of the Newfoundland Appalachians occurred in mid-Paleozoic times.

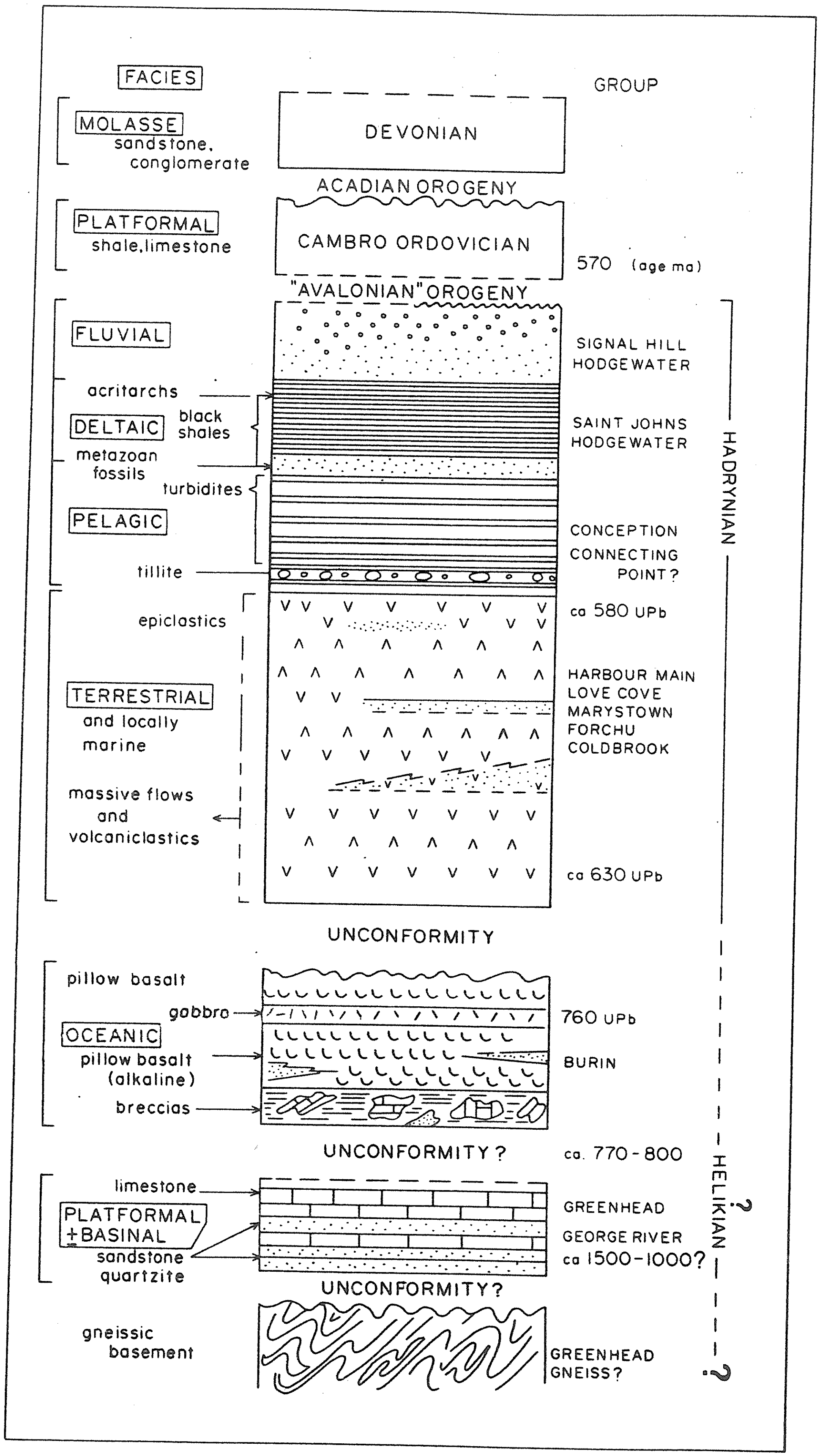


Figure 4: Schematic composite section of the Avalon Zone in Newfoundland and related rocks in the Canadian Appalachians

Little is known about the deep crust of the Avalon Zone in Newfoundland. The silicic nature of much of the widespread plutonic and volcanic activity typical of the zone, is suggestive of a major continental component to the Avalonian deep crust. Refraction data is sparse, but indicates that this crust may be partially transitional (between continental and oceanic) in nature. Few isotopic data from this region are availiable and the nature of this crust remains an outstanding problem. Correlations have been made between the Avalonian of Newfoundland and the latest Precambrian volcano-magmatic terranes of the Amoricain massif in the Channel Islands and Brittany, and with Northwest Africa; it is instructive to note that in each of the latter areas, the latest Precambrian rocks display genetic links to 2000 Ma gneissic basement (e.g. Roach and Topley, 1986; Kroner, 1980). Finally, deep seismic data has shown the Moho beneath the Avalon Zone to have a distinctive topography, offset by east-dipping reflectors. These may relate to deep crustal attenuation in the style of the Precambrian of the Pan-African belts (Kroner, 1980) or extensional tectonics in continental arcs (c.f. Wernicke, 1985).

The volcanic and plutonic rocks that are the hallmarks of the zone in Newfoundland and elsewhere appear to have formed in a short time span, compared to similar activity elsewhere in the orogen. Precise age dates suggest that the volcanism and plutonism may have occurred over a span of about 50 million years. Particularly distinctive aspects of this activity are the Avalonian granites. Voluminous calk-alkaline diorite - granodiorite - granite magmas were intruded, over a short time, into a region affected only by low grade metamorphism and a relatively simple structural style characteristic of the later pulses of the Avalonian orogen. The heat source for this magmatism is a fundamental problem with important ramifications for Avalonian tectonics.

The late Precambrian volcanic successions display a distinctive magmatic evolution and can be generally subdivided into three volcanologically, petrologically and geochemically distinct groups (O'Brien et al., 1986). The oldest is characterized by the eruption of subaqueous and submarine rhyolite, rhyodacite, basalt and, more rarely, andesite of calc-alkaline to mildly tholeiitic affinity and by the intrusion of calc-alkaline granodiorite and adamellite. A second interval is marked by extrusion of flood basalt and rhyolite of mild to strong alkaline affinity. A final interval, which may be continuous with the above alakaline volcanicity, is characterized by peralkaline and alkaline volcanic rocks and distinctive intrusive suites comprised of alkaline gabbroic rocks and peralkaline granites. This magmatism accompanied terrestrial sedimentation that occurred, at least locally, within fault-bounded basins that formed in response to strike-slip fault movements. The composition of the latest magmatic pulse is typical of post-Avalonian volcanicity within Newfoundland (O'Brien et al., 1983; O'Brien, unpublished data). The magmatic evolution of the Avalon is analogous to many of the latest Precambrian belts of Gondwana, particularly the Pan African of the Afro-Arabian region (e.g. Hijaz Arc). Modern anologues with similar evolutionary magmatic trends are found in continental rifts and ensialic arcs (e.g. Basin and Range, Patagonia).

The Precambrian sedimentary record of the Avalon Zone has an important bearing on tectonic models. The Avalonian sedimentary basins are dominant features of the zone, both onshore and offshore, and any model must explain their presence. They are characterized by extremely thick basin-fills, which record an evolution from complex packages of marine turbidites, through deltaic facies rocks, locally into alluvial deposits. The Avalonian Orogeny is recorded within the sedimentary succession in most areas by an angular unconformity; however, there is no apparent metamorphic break within the sequence, and in places, facies on either side of the unconformity can be genetically related. Locally, a genetic link between the sedimentary and volcanic rocks is preserved. In one such example (Connecting Point Group) the sediments display many characteristics of clastic facies found in volcanic arc basins. At present, it is not known whether the late Precambrian Avalonian sediments formed in one large basin produced by foundering of a volcano-plutonic substrate (c.f. the Lord Howe Rise; Jongsma and Mutter, 1978), or whether they occupied several distinct basins. In this respect, it is important to note that the distinct geophysical signature of the zone may reflect an original late Precambrian configuration of linked volcanic ridges and sedimentary basins, which is markedly similar to the Marianas region of the Pacific (e.g. Hussong and Uyeda, 1981).

DAY 1 (MORNING)

Tours of the surface and underground workings and of the mill are arranged. The stop descriptions given in the following account deal only with the surface workings.

Granite-hosted fluorspar deposits of the St. Lawrence area, Newfoundland

by

C.J. Collins

The following section includes a discussion of the geology of the St. Lawrence Granite and its associated fluorite deposits, the nature and genesis of the fluorite deposits and a description of the field trip stops in the St. Lawrence area.

Regional geology and mineralization of the St. Lawrence area

This description of the St. Lawrence Granite and its associated fluorite deposits is modified from Strong et al. (1978a), Howse et al. (1983), Collins (1984a) and Dickson (1984).

The St. Lawrence area, at the foot of the Burin Peninsula in eastern Newfoundland, hosts numerous fluorite-bearing veins which range up to 2 km in length and, in some cases, exceed 30 m in thickness.

Interest in fluorspar mining in the St. Lawrence area of Newfoundland began in 1928, with the first ore being extracted in 1933. The development of the mines was dominated by two companies, both of which eventually came under the control of ALCAN. In 1977, after 44 years of mining during which the St. Lawrence area was the only commercial producer of fluorite in Canada, ALCAN closed the mines. Mineral rights reverted to the province of Newfoundland and Labrador in 1983 and were subsequently granted to Minworth Ltd, which resumed production in 1986.

The general geology and location of the principal mineralized veins in the St. Lawrence area is shown in Figure 5.

The St. Lawrence Granite is a high level, peralkaline, Carboniferous pluton which intrudes volcanic and sedimentary rocks of Late Precambrian and Cambrian age. It shows a progressive decrease in grain size towards its contact. In some cases, separate minor, irregular phases with sharp intrusive contacts are found, with a generally younger age, from coarse to medium to fine grained to porphyritic. Miarolitic structures are seen within a few metres of the contacts. Tuffisites are common, consisting of discrete angular to rounded fragments of the granite in a finely comminuted matrix, and pink quartz-feldspar porphyry dikes related to the pluton are common outside its western margin (Teng, 1974; Teng and Strong, 1976).

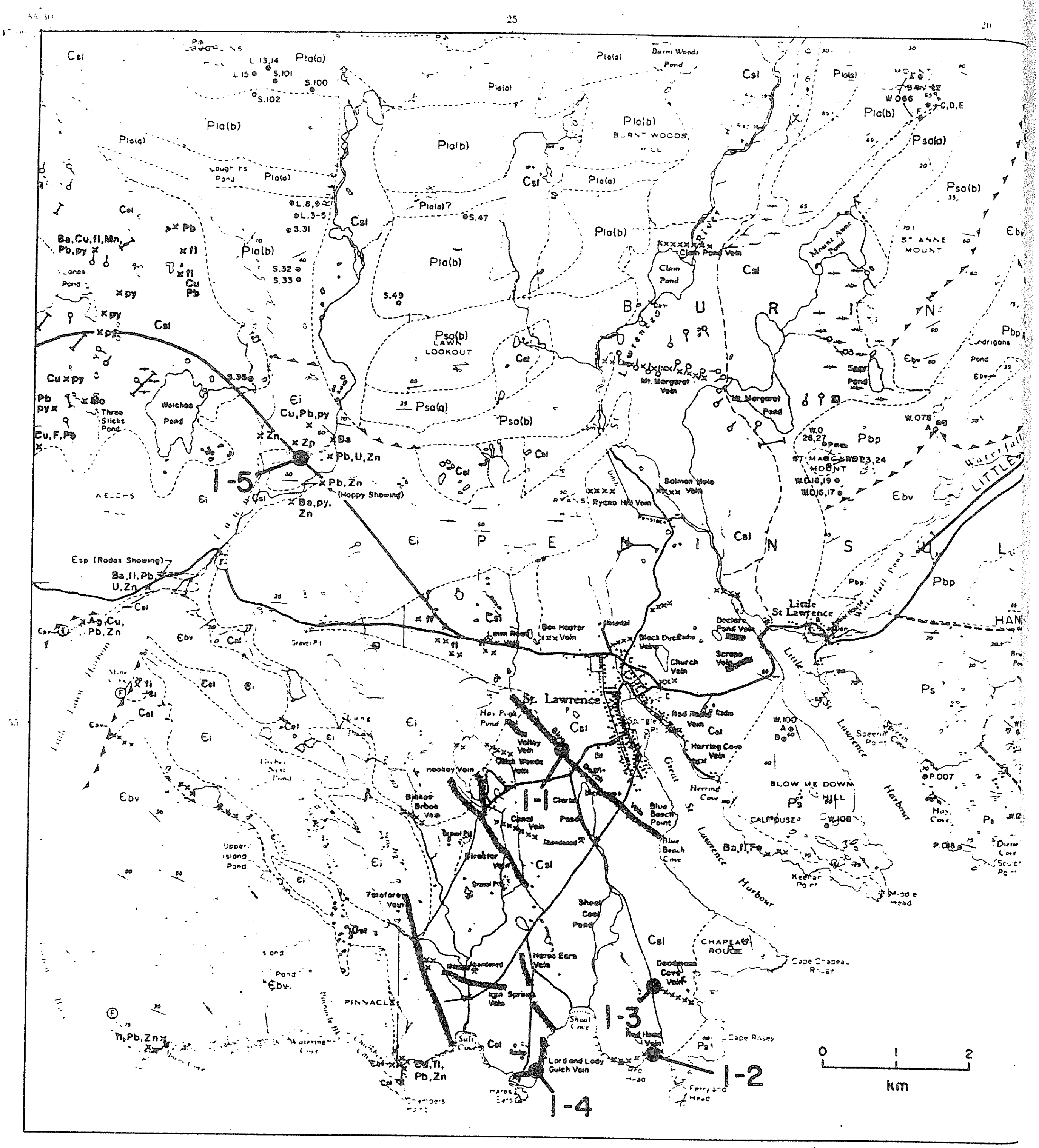
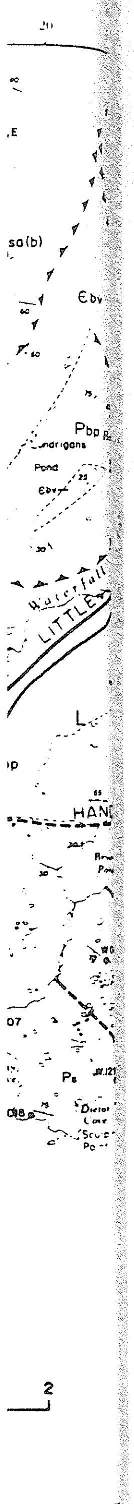


Figure 3 . Geology of the St. Lawrence area (Strong et al., 1978a).



The St. Lawrence Granite is a pink to red granitic rock consisting of red alkali feldspar and quartz, and characterized by a scarcity of ferromagnesian minerals. Its texture varies from medium grained hypidiomorphic to aphanitic, depending upon proximity to the contacts. The mafic minerals and plagioclase, where present, tend to be euhedral, whereas most of the alkali feldspar is subhedral; quartz occupies irregular interstices. Miarolitic cavities are commonly lined with quartz and, in some cases, fluorite crystals.

Petrographic studies show that the granite is composed essentially of quartz, orthoclase and albite, with minor amounts of riebeckite, aegirine, biotite, fluorite, magnetite and hematite (Teng, 1974). The granite contains 20 to 40% quartz, generally clustered between the larger feldspars, but near the pluton margins and in dikes, quartz forms embayed phenocrysts about 2 mm in length, as well as granophyric intergrowths. Alkali feldspar is the most abundant mineral, making up 30 to 60% of the rock. It is generally turbid with finely disseminated hematite, and perthitic textures. Albite rims are a common feature of the perthites. Modal plagioclase (albite) content of the St. Lawrence Granite ranges up to 20%, most of it exsolved from the alkali feldspars.

Riebeckite, where present, varies in grain size from 1 mm to 3 mm; it is generally associated with aegirine and is locally the more abundant of the two. Some thin sections contain up to 5% riebeckite, which appears to have crystallized after aegirine, as shown by riebeckite growing along the cleavages of aegirine crystals. Hornblende is present only in the marginal rocks from the west lobe of the pluton north of Lawn. In these areas the granite locally approaches granodiorite in composition, consisting of a small amount of quartz and a higher proportion of plagioclase (oligoclase) to orthoclase, combined with an increase in hornblende to about 10%, and sphene as an important accessory mineral. Some thin sections contain up to 10% (chloritized) biotite, but it is generally absent. Fluorite crystals are commonly associated with biotite. Hematite and magnetite are common accessory minerals. Zircon and apatite are rarely seen.

The "rhyolite porphyry" consists essentially of the same minerals as the St. Lawrence Granite, the only difference being in the texture. Feldspar phenocrysts are commonly perthitic orthoclase, but some are sanidine, and albite is relatively rare. Quartz is commonly euhedral, but shows embayed and corroded boundaries. Ferromagnesian minerals are not present as phenocrysts and are only rarely found in the groundmass. Chlorite is the only recognizable mafic mineral, and is secondary after biotite. Magnetite and hematite are disseminated in the groundmass with quartz and feldspar.

The St. Lawrence Granite and associated rocks are unique in eastern Newfoundland, in that they contain modal aegirine and riebeckite, peralkalinity, high concentrations of trace elements such as Rb, F, Nb and Zr and low K/Rb ratio, which result from advanced differentiation and volatile enrichment.

The high 87Sr/86Sr initial ratio of 0.722 (Bell and Blenkinsop, 1975) suggests substantial crustal involvement in their origin. Teng and Strong (1976) suggested an origin by crustal melting, and subsequent plagioclase fractionation from the parental melt.

The bulk compositions of St. Lawrence Granite plot near the ternary minimum at 50 MPa in the Q-Or-Ab-H_2O system, suggesting crystallization under very shallow conditions, in agreement with field evidence discussed above. These rocks are closely comparable to the Mesozoic Younger Granites of Nigeria which have been related by Black (1969) and numerous later authors to the disruption of Gondwanaland; the St. Lawrence Granite may have a similar, albeit earlier, relation to this disruption.

Fluorite Veins

About 40 distinct fluorite veins are known in the St. Lawrenc area, but not all of these are commercial deposits. Recent production has been from the Tarefare and Blue Beach veins. Almost all of the veins lie entirely within the granite or its related "rhyolite porphyry" dikes, and all are fissure fillings located in tension cracks that developed as a result of regional stress and contraction during cooling of the granite. The veins may vary from a few hundred metres to well over 2 km long and show a strong tendency to pinch and swell in both horizontal and vertical directions, producing a series of lenticular orebodies within each vein system.

Two major groups of fluorite veins are recognized:

(1) The low grade veins are generally widest, averaging 7 m, with CaF_2 content about 70%. Their strike varies from 135° to 160°. Examples are the Director, Tarefare and Blue Beach veins.

(2) The high grade veins have an average width of 1 m or less, and are characterized by acid grade ore averaging 95% CaF_2. Their strike varies from 60° to 115°. Examples are the Lord and Lady Gulch, Iron Springs and Canal veins.

The fluorite is mostly massive and coarsely crystalline, filling veins from wall to wall. Fluorite close to the vein walls is commonly finely banded, suggesting rhythmic deposition; this is especially characteristic of high grade veins.

Vugs and openings are commonly lined with exceptionally well formed crystals of fluorite, commonly as cubes which may reach 35 cm along an edge. Octahedral crystals appear to be restricted in occurrence to veins in "rhyolite porphyry" dikes and the country rocks, notably the Grebes Nest and Mine Cove veins.

The most common gangue mineral is quartz, occurring in the material locally known as 'blastonite'. The blastonite consists of brecciated fluorite cemented by a mixture of microcrystalline quartz and fluorite. The 'blastonite' was probably formed by both gas- and fault-brecciation.

Calcite occurs in nearly all the fluorite veins. It is common in nodular ores where it alternates with fluorite bands. Barite is a common gangue mineral; the Iron Springs and Canal veins have excellent honey-yellow tabular crystals intimately intergrown with the fluorite. At Lord and Lady Gulch, Meadow Woods, Blue Beach and other veins, barite occurs as white and pale red platy aggregates intergrown with fluorite.

Sulphide minerals are found in most veins in the St. Lawrence area. Galena and sphalerite are the most common, but chalcopyrite and pyrite have also been noted. Van Alstine (1948) reported covellite, bornite and chalcocite in several veins.

Hematite is a common accessory mineral in the fluorite and granite. It accounts for the color in all the red fluorite found in the area.

Genesis of the St. Lawrence Fluorspar Deposits

The first study of the genesis of the St. Lawrence deposits was made by Van Alstine (1948). He recognized a paragenetic sequence as shown in Figure 6. Fluorite, hematite, calcite, pyrite, chalcopyrite, sphalerite, galena, barite and quartz were all classified as primary "hypogene" minerals whereas chalcocite, covellite, malachite, azurite, chrysocolla, limonite and psilomelane were all classified as secondary "supergene" minerals. Hematite, calcite and quartz are found as both primary and secondary minerals. Van Alstine (1948) classified the St. Lawrence fluorite veins as epithermal, based on the low-temperature and low-pressure mineral assemblages, the vuggy nature of the veins and the abundance of breccia, colloform and crustification structures.

Van Alstine (1948) was also the first to point out a genetic relationship between the fluorite veins and the St. Lawrence Granite. He based this hypothesis on: (1) the close spatial relationship between the fluorite veins and the granitic rocks; (2) the abundance of fluorite as an accessory mineral in the granite; (3) the high concentration of fluorine in the granite; and (4) the presence of euhedral, colorless or purple fluorite cubes in the small vugs or miarolitic cavities within the granite. These factors led him to conclude that the source of the fluorite was in the same magma chamber that supplied the granitic magma and that the ore fluids were released as a result of late-stage differentiation, forming fluorite veins along faults and shears formed as the granite cooled and contracted. These conclusions have been supported by later studies (Williamson, 1956; Teng, 1974; Teng and Strong, 1976; Strong et al., 1984).

Williamson (1956) studied the structural patterns of the fluorite veins and the granite. He concluded that they reflect the earlier tectonic history of the underlying rock units. He recognized the cyclic nature of the mineralization and related it to repeated movement along the veins, creating new spaces for ore deposition and causing brecciation of pre-existing vein material and country rock. He pointed out that the lack of fluorite mineralization in the overlying sediments was due to the fact that open fissures narrowed and closed on passing from the granite into the country rock. Williamson (1956) concluded that the primary control on fluorite mineralization was the development of a fracture pattern due to a stress system produced by the cooling of the granite, but whose orientation was dictated by an original stress system developed in the underlying rocks by earlier tectonic forces. He also attributed the orientation of the elongate granite body to earlier tectonic controls.

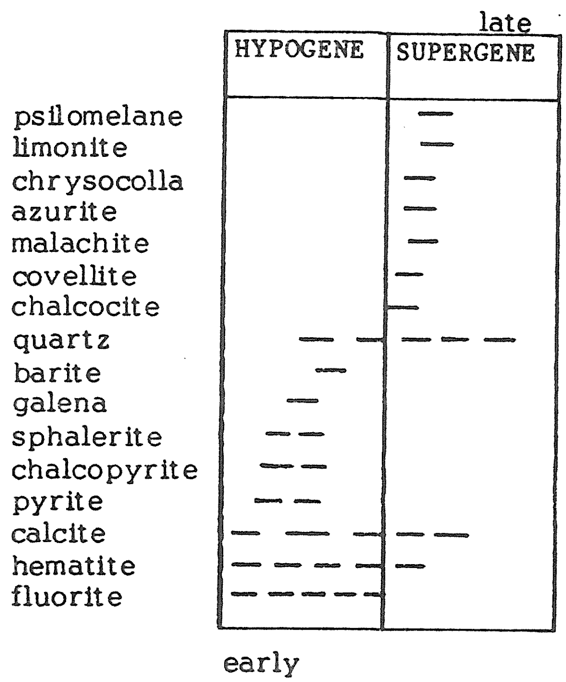


Figure 6 . Paragenesis of the St. Lawrence fluorspar veins (Van Alstine, 1948).

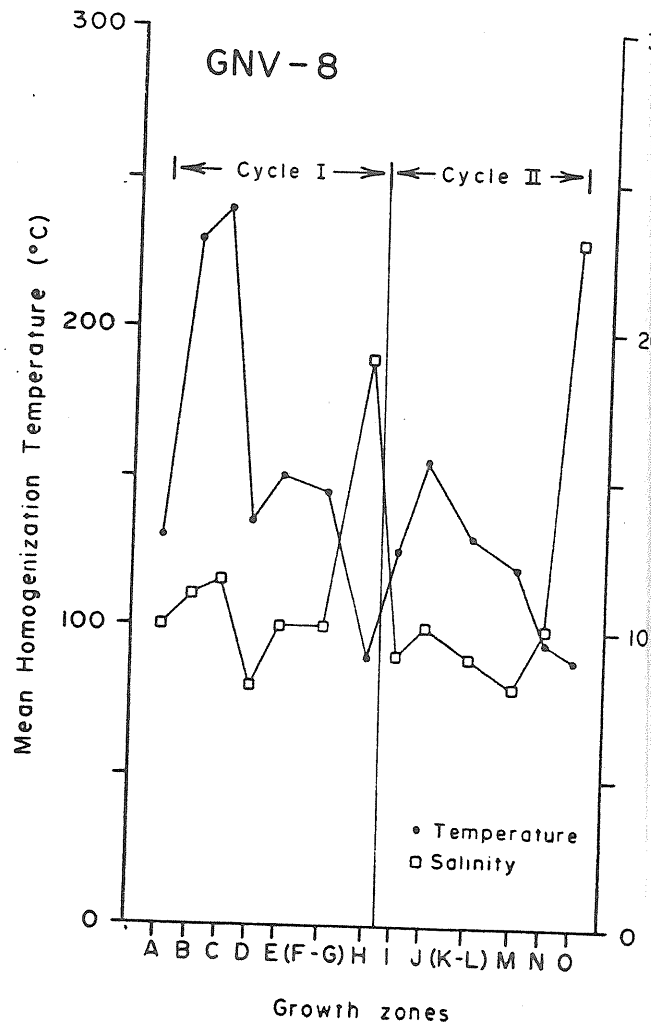
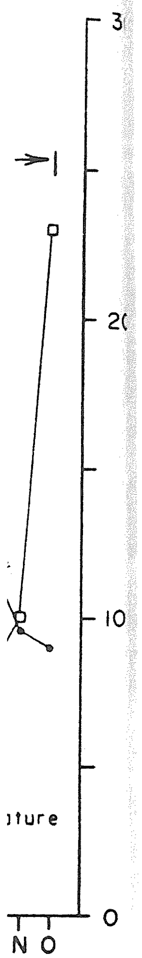


Figure 7 . Plot of salinity and homogenization temperature for the Grebes Nest Vein (GNV-8).

Teng (1974) supported many of the views of Van Alstine (1948) and Williamson (1956) regarding the genesis of the fluorite veins. He added that continued movement along the fractures produced varying degrees of brecciation of the wall rocks. Later volatile materials were separated and escaped along innumerable channelways to be deposited as veins in the structure already developed in the upper part of the granite. The escape of volatiles beyond the boundaries of the granite was largely prevented by the impermeability of overlying hornfels which effectively impounded them. Such a 'ponding' of fluids, at the granite contact, can result in alteration of the granite such as that described by Strong (1982).

Based on their own results and the experimental results of others, over a wide range of salinities, Richardson and Holland (1979b) suggested that fluorite is deposited as a result of: (1) decreasing temperature and pressure along the path of fluid flow, (2) fluid mixing to cause dilution or compositional change, or (3) fluid interaction with wall rocks to cause an increase in pH. The conclusion of a study by Strong et al. (1984) is that all but one of the specific mechanisms of fluorite precipitation proposed by Richardson and Holland (1979a, b) can be eliminated for the St. Lawrence veins. The results of fluid inclusion studies, oxygen-isotope studies and REE geochemistry all appear to rule out precipitation of fluorite as a result of a decrease in temperature, wall rock alteration or fluid mixing. They emphasize an increase in pH as an important potential cause of fluorite deposition. They suggest that the boiling of fluids, and in particular the ebullition of carbon dioxide, possibly along with a loss in fluorine as SiF_4 or HF, would provide a ready means of increasing the pH, and thus exercise the main control on fluorite deposition.

A thesis study by Collins (1984a) supports the findings of Strong et al. (1984) that boiling of fluids, causing an increase in pH, is a major control in fluorite deposition, but not necessarily the only one. From the study of fluid inclusion data, Collins (1984a, b) suggests that fluorite deposition is cyclic, each cycle being characterized by an abrupt increase in temperature and salinity followed by a more gradual decrease in temperature and salinity. Figure 7 shows this cyclic nature of deposition in the Grebes Nest Vein. The end of each cycle is characterized by a low temperature - high salinity fluid. This implies a mixing of cool, high saline connate waters with cooling magmatic fluids, or it may represent the residual liquid remaining after boiling has occurred. The escape of low salinity vapor would cause an increase in salinity of the residual liquid, and subsequent rise through cooling country rock and previously deposited fluorite would effectively cool the fluid. Rare inclusions providing extremely low salinity values (around 2 eq.wt% NaCl) could represent condensed vapor which boiled off from the fluids to condense later in cooler regions of the conduit system.



1
erature
Vein

Field Stops

Five field trip stops are planned for the St. Lawrence area. The locations of these stops are shown on Figure 5.

STOP 1-1: Blue Beach North Deposit

This deposit is located immediately west of the town of St. Lawrence and is accessible by road. The vein is well exposed at the site of the open cut operations. This deposit is currently mined by open cut while underground development prepares the deposit for underground mining. The vein is hosted by the St. Lawrence Granite except in its northern extremities where it rapidly pinches out as it enters the overlying metasediments of the Inlet Group. The Blue Beach vein strikes approximately 135° and has a varible dip which is generally steep to the west, but dip reversals are common with depth and along strike. The Blue Beach vein structure has an overall strike-length of 2350 metres with widths ranging from 0.5 metres to more than 20 metres.

The Blue Beach Vein is hosted almost entirely by the St. Lawrence Granite except in its northern and southern extremities. To the north, the vein terminates where it intersects a shallowly dipping (northward) cap of metamorphosed sediments of the Inlet Group. To the south, the vein pinches out in the granite near the contact with metavolcanics of the Burin Group. The St. Lawrence Granite, in the vicinity of the vein, is predominantly medium to coarse grained pink granite with phenocrysts of quartz and orthoclase. Tuffisites are common, especially at or near vein contacts.

The Blue Beach North Deposit contains a variety of ore types ranging from coarsely crystalline fluorite to low grade granite breccia. The coarsely crystalline ore is primarily fluorite although minor zones of calcite and quartz occur. The breccia ores have a variety of mineralogical types with the matrix being any combination of quartz, fluorite or calcite and fragments being any combination of quartz, fluorite, calcite, breccia ore, granite or tuffisite. The granite breccia ore consists of a high percentage of granite fragments cemented by a coarsely crystalline fluorite matrix. The vein structure contains zones of "sand" or "pug" which have been found to be primarily fine to medium sized fragments of fluorite-rich ore. These zones are believed to be equivalents of the breccia-type ores which were formed by late movements and never cemented.

The main gangue minerals are quartz and calcite which occur in a variety of textures as mentioned above. Barite is common in certain parts of the vein system and usually occurs as coarsely crystalline pink to white zones. Sulphide minerals are common especially galena and sphalerite with lesser amounts of chalcopyrite. Hematite is common and is evidenced by distinct zones of red-brown ore. Another oddity of the Blue Beach Vein is the presence of "mud" seams. These are a black mixture of sandy ore material with a high percentage of black manganese-rich material. It has been suggested (Williamson, 1956) that these seams were formed as a result of groundwater percolating downward through the vein material. These "mud" seams have been found in mine workings as deep as 100 metres.

Stop 1-2: Red Head Vein

This vein is located approximately 3 km south of St. Lawrence and is accessible by road. The fluorite vein occurs near the coast and is well exposed. The vein is hosted by the St. Lawrence Granite. The Red Head Vein strikes 080° and has a variable dip ranging from 70°S to 80° N. The vein has a known strike length of 550 m along which its width pinches and swells from 0.7 to 3 m.

The Red Head Vein occurs within the St. Lawrence Granite near its contact with Burin Group sediments to the south. Here the granite is medium to coarse grained pink alaskite. Tuffisites, consisting of discrete angular to rounded fragments of the granite in a finely comminuted matrix, can be found at this locality.

The Red Head Vein consists of coarsely crystalline fluorite which usually fills the vein from wall to wall. On the western part of the vein it is characterized by 'breccia ore' and offshoot veins. The 'breccia ore' consists of angular to subrounded fragments of granite cemented in a matrix of coarse crystalline fluorite and may have a simiar origin as the tuffisites. The fluorite is coarsely crystalline and varies in color. Typical colors are shades of pink, white, green and purple. Sulfide minerals can be found in this vein, especially galena and sphalerite.

STOP 1-3: Rosy Ridge Vein (Deadman's Cove Vein)

This vein is located approximately 500 m north of the Red Head Vein (Stop 1-2). The vein is well exposed in an open cut working as well as in a series of pits toward the coast. The Rosy Ridge Vein strikes approximaely 110°and has a variable dip which is close to vertical. The vein has a known strike length of approximately 700 metres and a width which pinches and swells, averaging 1 metre.

The Rosy Ridge Vein is hosted by the St. Lawrence Granite along its entire length. The western end terminates in the granite whereas the eastern end terminates near the contact of the St. Lawrence Granite with the Burin Group metavolcanics. The granite here is similar to that of the Red Head Vein.

The mineralogy of the Rosy Ridge Vein is also similar to that of the Red Head Vein. The ore is primarily coarsely crystalline fluorite, but well developed breccia zones can be seen locally. Sulphide minerals are present but not as abundant as at Red Head.

STOP 1-4: The Lord and Lady Gulch Vein

This vein is located near the coast approximately 5 km southwest of St. Lawrence and is accessible by road. It is hosted by the St. Lawrence Granite and is well exposed in old mine workings at its eastern end. The vein strikes 070o and dips 80oS and has a strike length of 365 m. It averages 2.5 m wide and locally attains widths up to 6.5 m. The vein is known to pinch and swell both laterally and vertically, producing lens-shaped bodies.

The Lord and Lady Gulch Vein consists of coarsely crystalline fluorite which fills the vein from wall to wall. Wall contacts are well defined but commonly intensely brecciated. Slickenides can be observed on some wall contacts. Although the vein is continuous over 365 m, it does have a tendency to develop branch veins. The fluorite mineralization occurs in a variety of colors including shades of red, pink, yellow, white, green and purple. Barite is found in this vein as well as galena and chalcopyrite. Minor calcite is the only gangue mineral.

STOP 1-5: The Granite Contact (Metasomatism)

The contact of the St. Lawrence Granite with its country rock is generally sharp with little or no alteration of the granite, reflecting the generally low activity of water during deposition (Teng and Strong, 1976; Evans, 1977). In areas where the contact is perpendicular to the bedding or fractures, recrystalliation, alteration and deposition of fluorite and other minerals permeate tens of metres into the country rocks. Where no such permeability exists, the fluids appear to have been ponded within the pluton, beneath the contact, the main effect on country rocks being a zone of recrystallization, approximately 2 m wide (Strong, 1982). One such contact is exposed in a roadcut approximately 7 km west of St. Lawrence.

Here approximately 2 m of gently dipping shale forms a roof to the granite exposed in the lower 3 m of the outcrop. The intrusive contact is sharp with local brecciation and veining along it. The granite exposed in the roadcut represents a chilled margin of the pluton with phenocrysts of quartz and K-feldspar. Within 2 m of the contact, the granite is altered, with the most intense alteration occurring close to the contact and along fractures within the granite.

The alteration is progressive from unaltered granite 3 m from the contact to 'episyenite' at the contact (Strong, 1982). The least altered granite (3 m below the contact contains equal proportions of quartz and K-feldspar phenocrysts in a slightly oxidized, pink, fine grained groundmass. Near the contact (2 to 3 m below), there is a slight fading of the pink color due to albitization of the K-feldspars and an increasing proportion of quartz microveinlets. Within 2 m of the contact the alteration of the granite has produced an 'episyenite' which is marked by an abrupt disappearance of quartz and K-feldspar and an increase in calcite and albite (Strong, 1982). No further changes in major mineralogy toward the contact can be observed except for slight variations in proportions of calcite and albite and the presence of pyrite in gas breccia (a later event than the pervasive alteration). Small red sphalerite crystals and minor galena and fluorite can be seen within the alteration zone.

DAY 1: AFTERNOON
PRECAMBRIAN-CAMBRIAN BOUNDARY:
A CANDIDATE STRATOTYPE

(follow Route 220 to Grand Bank)

INTRODUCTION

The Marystown Group volcanics are overlain disconformably by a continuous sequence of siliciclastic rocks that span the Precambrian-Cambrian boundary: in ascending order, the Rencontre, Chapel Island and Random formations. The redbed conglomerates, sandstones and shales of the Rencontre Formation were deposited in a variety of fluvial and marginal marine settings (Smith and Hiscott, 1984). The sandstones, siltstones and minor limestones of the Chapel Island Formation were deposited in nearshore and shelf environments (Crimes and Anderson, 1985; Myrow, 1987). The sandstones, shales and quartzites of the Random Formation were deposited in tidally dominated nearshore and shoreline environments (Anderson, 1981; Hiscott, 1982). The Random Formation is stratigraphically the first regionally correlatable unit in the Newfoundland Avalon Zone (Fig. 8).

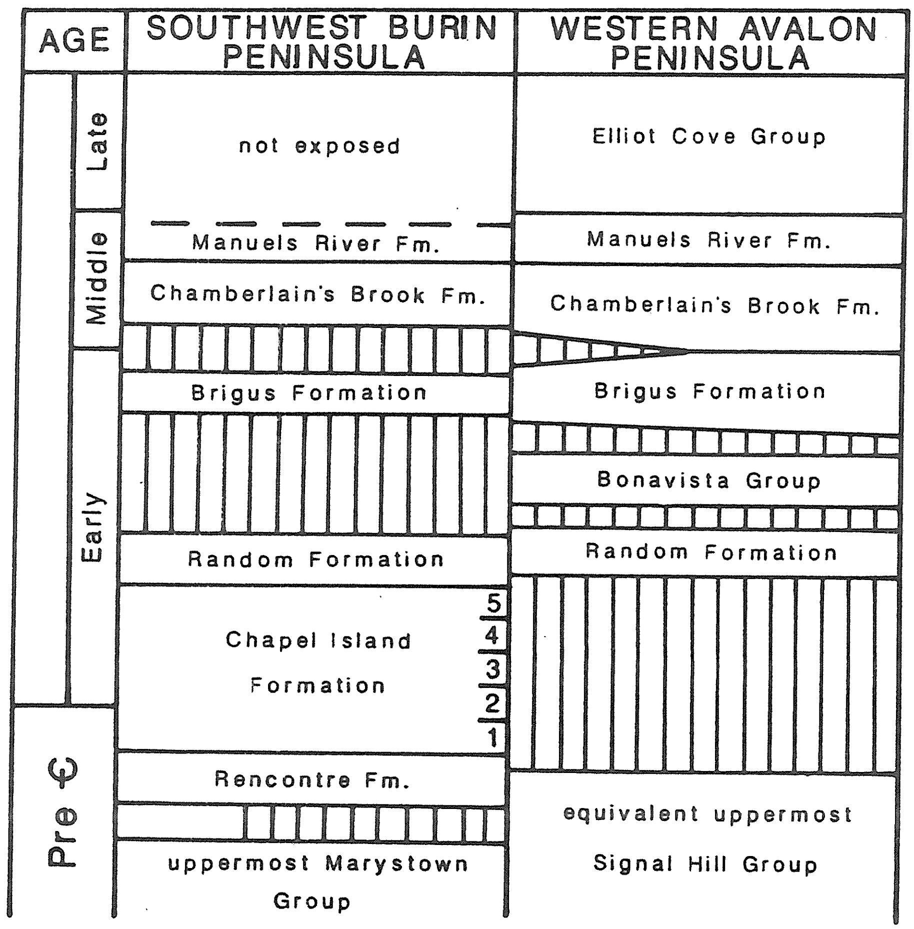


Fig. 8. Latest Precambrian to Cambrian correlations from Burin Peninsula to the western Avalon Peninsula (from King, 1980; Strong, 1979); only the Signal Hill Group is present in this time interval on the eastern Avalon Peninsula. Vertical rule indicates major unconformities.

In the Fortune Bay area, the Random Formation is in conformable contact with the underlying Chapel Island Formation, whereas elsewhere, to the northeast and east, its base is unconformable or the Random is absent (Hutchinson, 1953; McCartney, 1967; Fletcher, 1972; Anderson, 1981). Where the base is unconformable, the underlying strata were locally gently folded and deeply eroded prior to deposition of the Random Formation. Thus 1800 m of Precambrian sediments were eroded from an area southeast of Trinity Bay (McCartney, 1967), and over 450 m of Precambrian sediments were removed from the southeastern part of the Cape St. Mary's Peninsula (Fletcher, 1972). The Random Formation was deposited during a marine transgression, with the marine basin apparently opening toward the southwest (Anderson, 1981; Hiscott, 1982).

The Early Cambrian was a time of eustatic sea-level rise (Brasier, 1980; Mathews and Cowie, 1979; Vail et al., 1977). Anderson (1981, p. 825) considered the stratigraphic transition from the Rencontre Formation to the Chapel Island Formation to represent the initial stage of this global transgression. Given the complex history of changes in relative sea level evident in both the Rencontre Formation (Smith and Hiscott, 1984) and the Chapel Island Formation (Myrow, 1987), however, it seems unlikely that the influence of eustatic sea-level rise can be filtered out of the complexity of basin tectonics and changing depositional systems so as to clearly identify the initial stage of the Cambrian transgression in Newfoundland. However, the regional onlap displayed by the Random Formation and other quartzites of this age throughout the world does suggest deposition during a eustatic rise.

UPPERMOST PRECAMBRIAN-CAMBRIAN OF THE FORTUNE BAY AREA

Rencontre Formation

Coarse conglomeratic deposits of the Rencontre Formation overlie the acidic volcanics of the Marystown Group. These proximal alluvial deposits are succeeded by fluvial and marginal marine siliciclastic sandstones, siltstones and mudstones. Depositional environments for the upper part of the formation near the transition into the overlying Chapel Island Formation were probably peritidal (Smith and Hiscott, 1984; Myrow, 1987). Smith and Hiscott (1984) concluded that the Rencontre Formation was probably deposited during a period of active strike-slip faulting. Sedimentation took place in a narrow but expanding, fault-bounded, northeast-southwest trending basin.

Chapel Island Formation

Lithostratigraphy

The thickness of the Chapel Island Formation in the southwestern part of the Burin Peninsula is ca. 1000 m. Bengtson and Fletcher (1983) divided the formation into five informal members, numbered 1-5, based on the outcrops at Grand Bank (lower members) and Little Dantzic Cove (upper members). This same scheme was also used without modification by Crimes and Anderson (1985), who studied the same outcrops. The first detailed measured sections of the formation were prepared by Myrow (1987), who corrected errors in the thickness of members and better defined the member boundaries. The redefined stratigraphy is summarized in Figure 9; facies are described in Table 1. Correlations between outcrops of the Chapel Island Formation on the Burin Peninsula and within Fortune Bay (Fig. 10), including the positions of member boundaries, are given in Figures 11 and 12 (from Myrow, 1987; also see Narbonne et al., 1987).

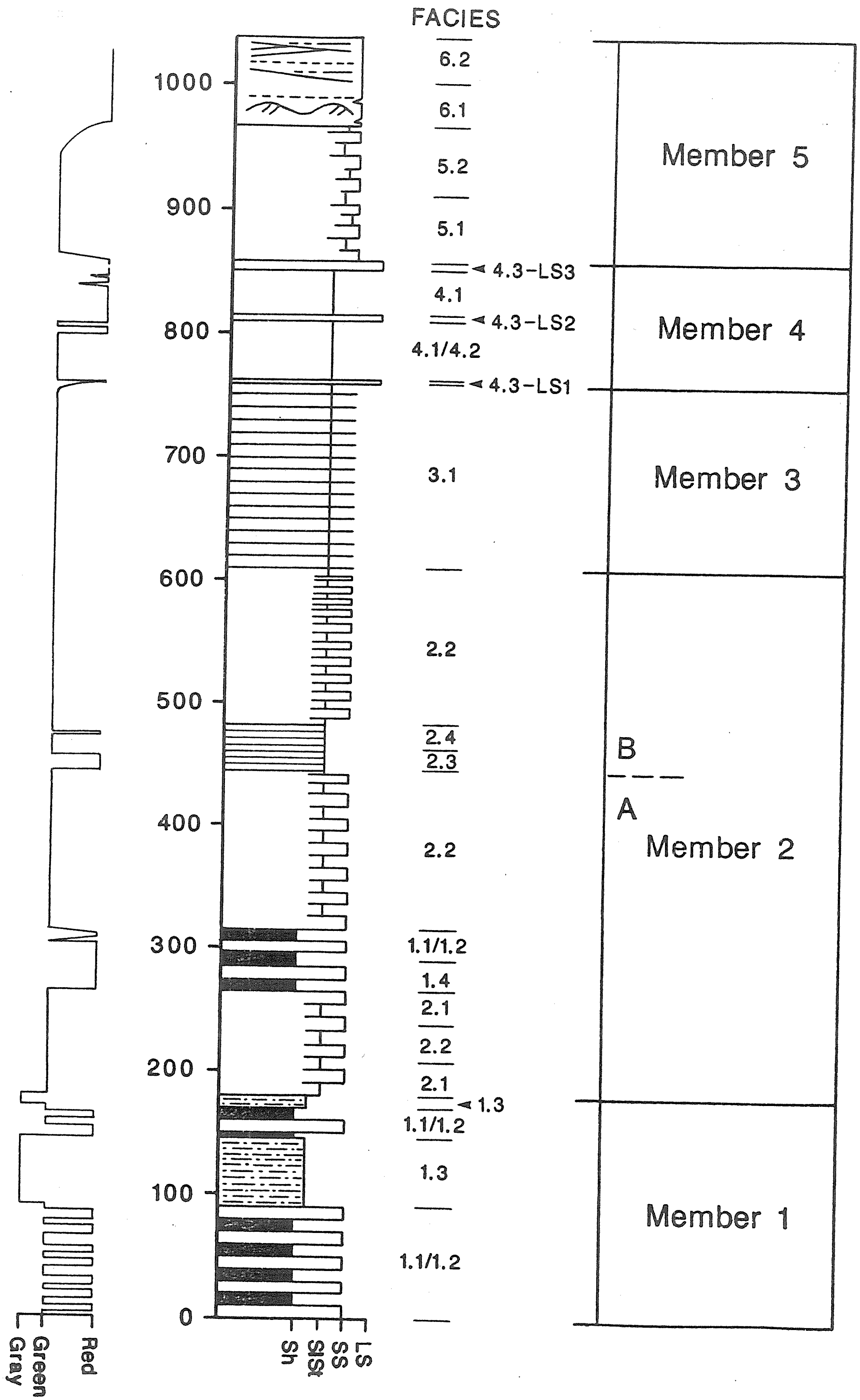


Fig.9 . Summary of stratigraphy of the Chapel Island Formation used in this guide. Colour and facies (Table 1) are indicated.

Table 1. Facies characteristics for Chapel Island Formation

FACIES	LITHOLOGY	SEDIMENTARY STRUCTURES	ENVIRONMENTAL INTERPRETATION
FACIES 6.1a/b	Red micaceous sandstone; minor siltstone and conglomerate	Parallel lamination; hummocky and swaly cross-stratification; wave ripples	Prograding storm-dominated shelf (nearshore)
FACIES 5.1/5.2	Interbedded green sandstone and sandy siltstone	Graded sandstones,; parallel lamination; current ripples; rare convolute bedding	Prograding storm-dominated shelf (offshore)
FACIES 4.3	Red to white micritic and stromatolitic limestone	Fossiliferous; desiccation cracks; tepee and sheetcrack structures; mud mounds; planar/columnar stromatolites; oncolites	Peritidal; low energy
FACIES 4.2	Gray mudstone	Burrowed; pyrite nodules; pyritic steinkerns	Low energy dysaerobic shelf
FACIES 4.1	Red and green mudstone	Bioturbated; abundant carbonate concretions	Low energy oxygenated inner shelf
FACIES 3.1	Green laminated siltstone	Parallel lamination; current ripples; carbonate concretions; current/parting lineations	Outer shelf (sub wave-base)
FACIES 2.3/2.4	Red and green laminated siltstone	Red: parallel lamination, wave-ripple lamination Green: carbonate nodules, pyrite nodules	Low energy subtidal shelf; Delta abandonment facies
FACIES 2.2	Thin to medium bedded gray-green sandstone and siltstone	Grading; wave and combined-flow ripples; pebble lags; parallel lamination; mass movement deposits; HCS; graded rhythmites	Storm-influenced subtidal Lower delta front/prodelta
FACIES 2.1	Very thin to thin bedded gray-green sandstone and siltstone	Abundant gutter and pot casts; pinch-and-swell and lenticular bedding; unifite siltstone beds	Storm-influenced shallow subtidal Upper shoreface/delta front
FACIES 1.4	Medium bedded red sandstone and shale	Channel sandstones; parallel lamination; shale rip-up clasts; synaeresis/desiccation cracks	Peritidal, high energy,tidal influence
FACIES 1.3	Gray to black laminated sandstone and shale	Contorted bedding; synaeresis cracks; pyrite and phosphate nodules; channel sandstones	Semi-restricted low-energy
FACIES 1.1/1.2	Red and green sandstone and shale	Flaser, wavy and lenticular bedding; scour and channel sandstones; synaeresis/desiccation cracks; parallel and ripple cross-lamination	Peritidal, tidally-influenced

The base of the Chapel Island Formation is most easily defined at Grand Bank Head where following Potter's (1949) suggestion it is arbitrarily placed at the first green interval greater than 1 m thick in the transition from the underlying redbeds of the Rencontre Formation. Member 1 consists of approximately 180 m of red and green, thin to medium bedded sandstones, siltstones and shales with intervals of laminated black shales and gray shaly siltstones in the upper half.

The contact between member 1 and member 2 is best exposed at Fortune Dump where it is placed at the lowest stratigraphic occurrence of the very diagnostic and easily recognizable gray-green siltstone and sandstone beds that characterize most of member 2. Packages of red and green sandstones and shales, similar in character to those in member 1, are present above this base in member 2.

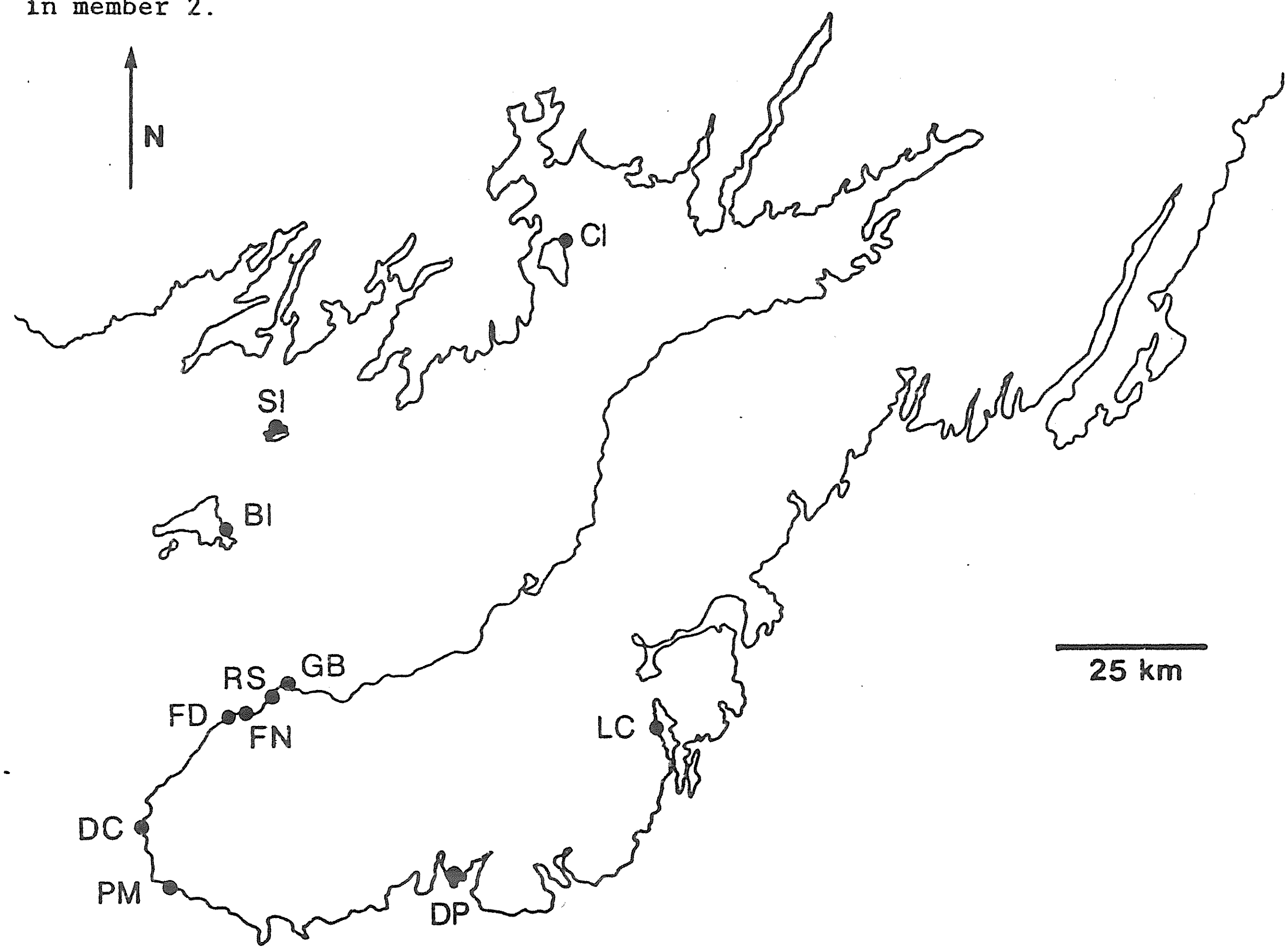


Fig. 10. Location of sections (Figs 6 & 7) in the Chapel Island Formation: BI, Brunette Island; CI, Chapel Island; DP, Duck Point; FD, Fortune Dump; FN, Fortune North; GB, Grand Bank Head; LC, Lewin's Cove; DC, Dantzic Cove; PM, Point May; RS, Radio Station; SI, Sagona Island.

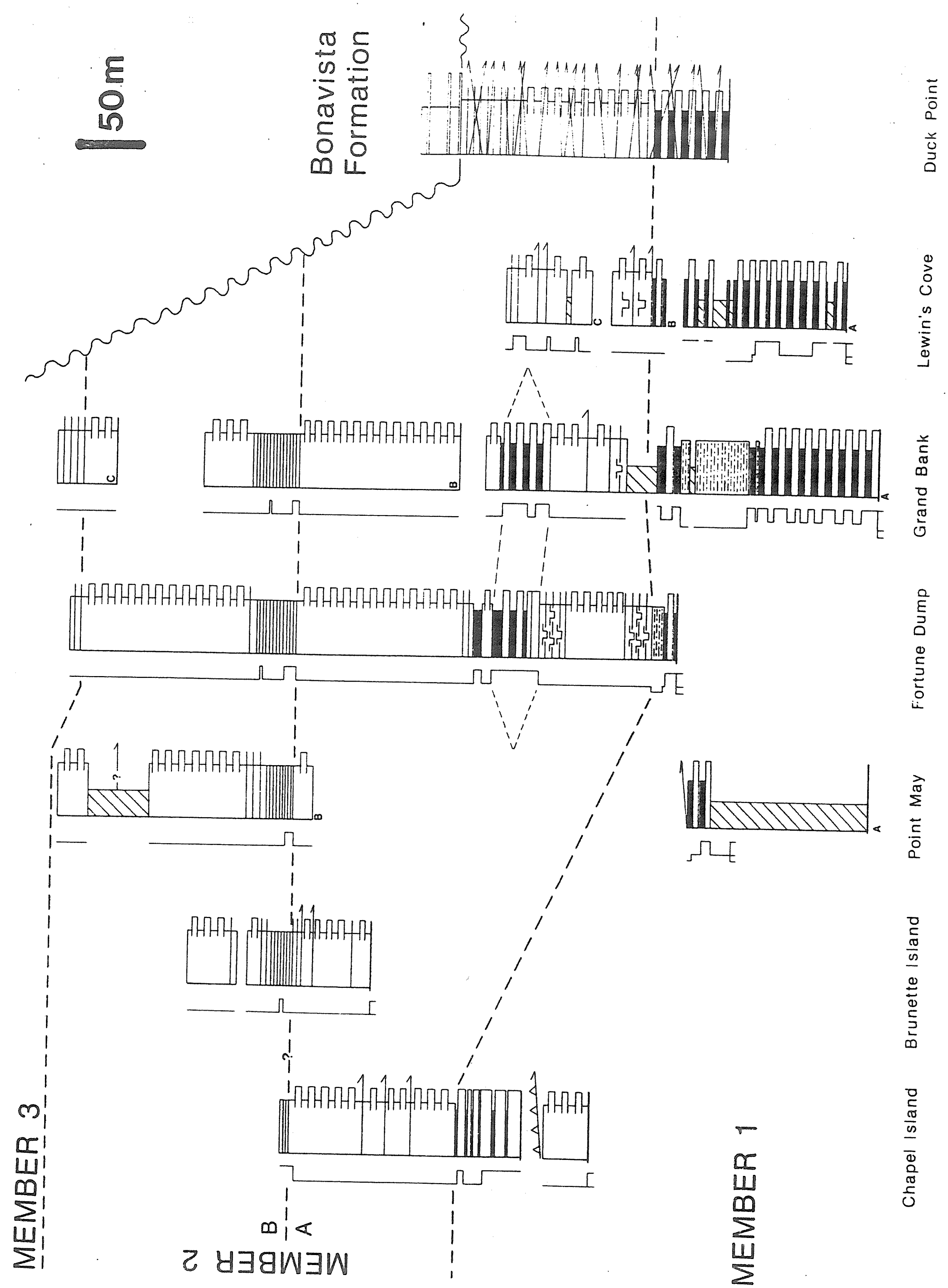


Fig. 11 . Correlation of the lower part of the Chapel Island Formation at Figure 5 localities (Myrow, 1987).

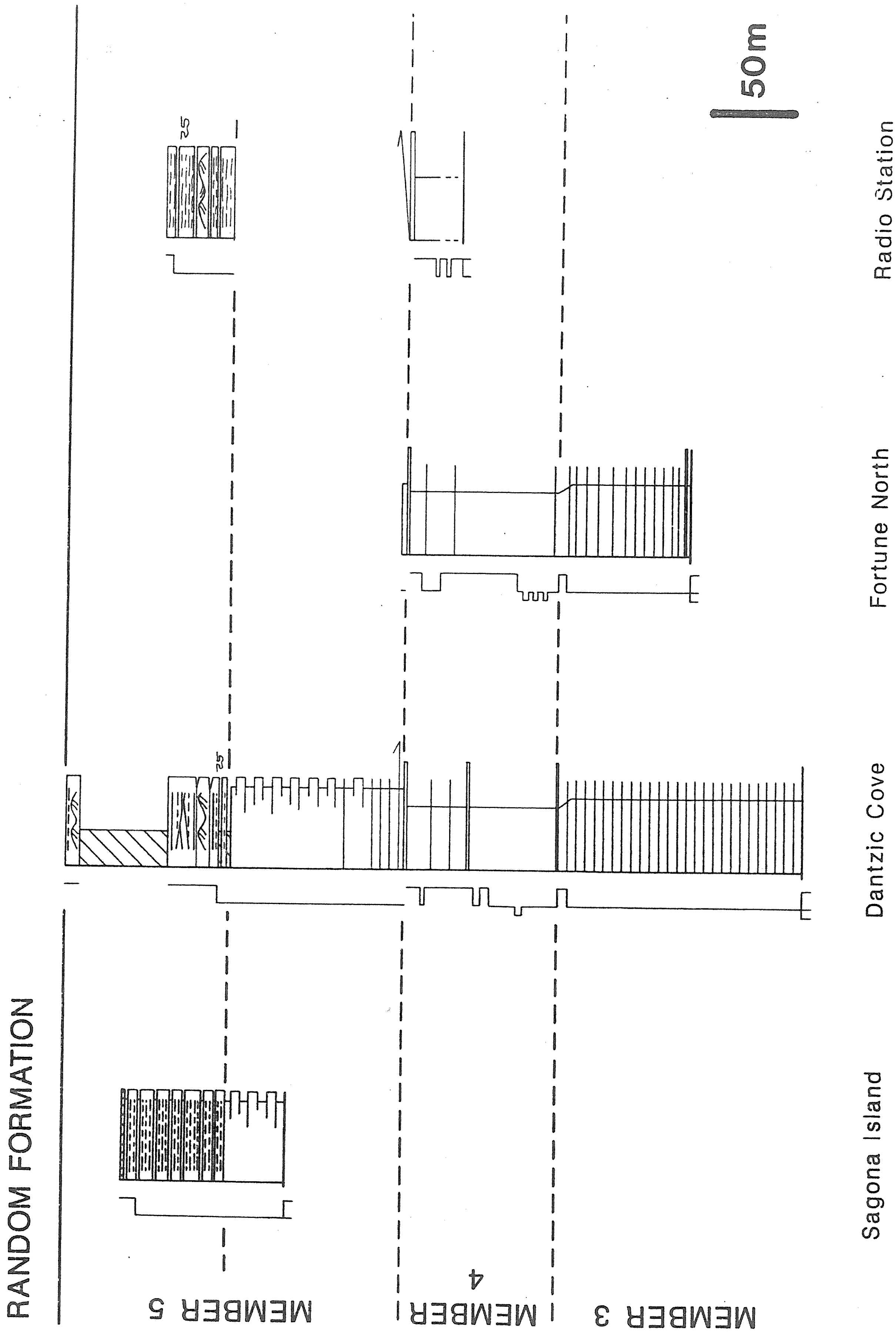


Fig. 12. Correlation of the upper part of the Chapel Island Formation at Figure 5 localities (Myrow, 1987).

The contact between members 2 and 3 was placed by Bengtson and Fletcher (1983) and Crimes and Anderson (1985) at the base of a red siltstone unit at Grand Bank Head (Facies 2.3 at 450 m, Fig.9). At this locality, the red siltstone is succeeded by several meters of green laminated siltstone (Facies 2.4) that occur just below the highly faulted part of this section. These earlier studies equated these laminated siltstones with member 3, which is lithologically similar and is found (with its base not exposed) at Little Dantzic Cove. At Fortune Dump, however, these red and green siltstones (formerly part of member 3) are succeeded by green sandstones and siltstones identical to those of member 2; this can also be demonstrated at Grand Bank Head by carefully tracing beds across numerous faults along the coast. The lower boundary of member 3 has therefore been moved upward from its former position to a point higher in the formation where the sandstones and siltstones grade into the laminated siltstones of member 3. The old boundary between member 2 and member 3, which now lies <u>within</u> member 2, is retained for historical reasons and separates the member into two parts, 2A and 2B. The early estimate of 50 m for member 2 has been revised to approximately 430 m: 265 m for 2A, and ca. 165 m for 2B.

The change from member 2 to the carbonate-concretion-bearing, laminated siltstones of member 3 is gradational, recording a decrease in sandstone percentage and bed thickness. The most complete stratigraphic section of member 3 is found at Dantzic Cove, where 135 m of strata are present. This is a minimum thickness for the member. Facies characteristics suggest that the thickness of member 3 is not significantly greater than the estimate of 150 m given by Bengtson and Fletcher (1983).

Member 4 consists of red and green bioturbated mudstones, gray pyritiferous mudstones, and thin, nodular to bedded, micritic and stromatolitic limestones. The contact between member 3 and member 4 is defined as the base of the first continuous limestone bed at Dantzic Cove (LS 1). A few meters of red mudstone underlie this first limestone and mark the top of the second of two, meter-scale, green to red mudstone cycles at the top of member 3. These cycles can be traced to Fortune North (Fig.10) where the red mudstones at the top of the second cycle define the top of member 3 in the absence of LS 1. The top of member 4 at all localities in the southwestern part of the Burin Peninsula is a 60-80 cm-thick red and white limestone with prominent algal stromatolites and mud mounds. The thickness of member 4 at Little Dantzic Cove and Fortune North is 85 m. The estimate of 165 m given by Bengtson and Fletcher (1983) was based on an incorrect correlation between a limestone bed (coalesced carbonate concretions) <u>within</u> member 3 at Fortune North and the lowest limestone bed (LS 1) at the <u>top</u> of member 3 at Dantzic Cove (T. P. Fletcher, pers. comm., 1985).

Member 5 overlies the thick stromatolitic limestone at the top of member 4 and consists, at the base, of green, thin to medium bedded sandstones and sandy siltstones that show an upward increase in bed thickness and sandstone/siltstone ratio. This trend continues into the upper part of member 5 which is dominated by medium to thick bedded, red micaceous sandstones. The thickness of member 5 at Dantzic Cove, from the top of member 4 to the lowest quartzose sandstone bed of the Random Formation, is 178 m.

Member 1 Sedimentology

Three lithofacies comprise member 1: Red and Green Sandstone/Shale Facies 1.1, Shaly Facies 1.2, and Gray to Black Shale Facies 1.3. The first two facies dominate the lower part of member 1 and represent (a) lower tidal flat/shallow subtidal and (b) middle/upper tidal flat deposits, respectively. Bedding characteristics, sedimentary structures and paleocurrent data (Fig.13) indicate that reversing tidal currents were the main agents of transport and deposition. Facies 1.3 was deposited in semi-restricted shoreline environments of uncertain affinity. Partially enclosed shoreline embayments including interdistributary bay environments, are considered probable settings for deposition of this facies.

Member 2 Sedimentology

Five lithofacies comprise member 2: Gutter Cast Facies 2.1, Sandstone/Siltstone Facies 2.2, Thinly Bedded Red Siltstone/Shale Facies 2.3, Green Laminated Siltstone Facies 2.4 and Medium-Thick Bedded Red Sandstone/Shale Facies 1.4. Facies 2.1 and 2.2 are considered upper delta front and lower delta front/prodelta deposits, respectively. Facies 2.3 and 2.4 are delta abandonment facies, and as such, form laterally extensive marker horizons within the central part of member 2. Facies 1.4 represents fluvial/marginal-marine environments.

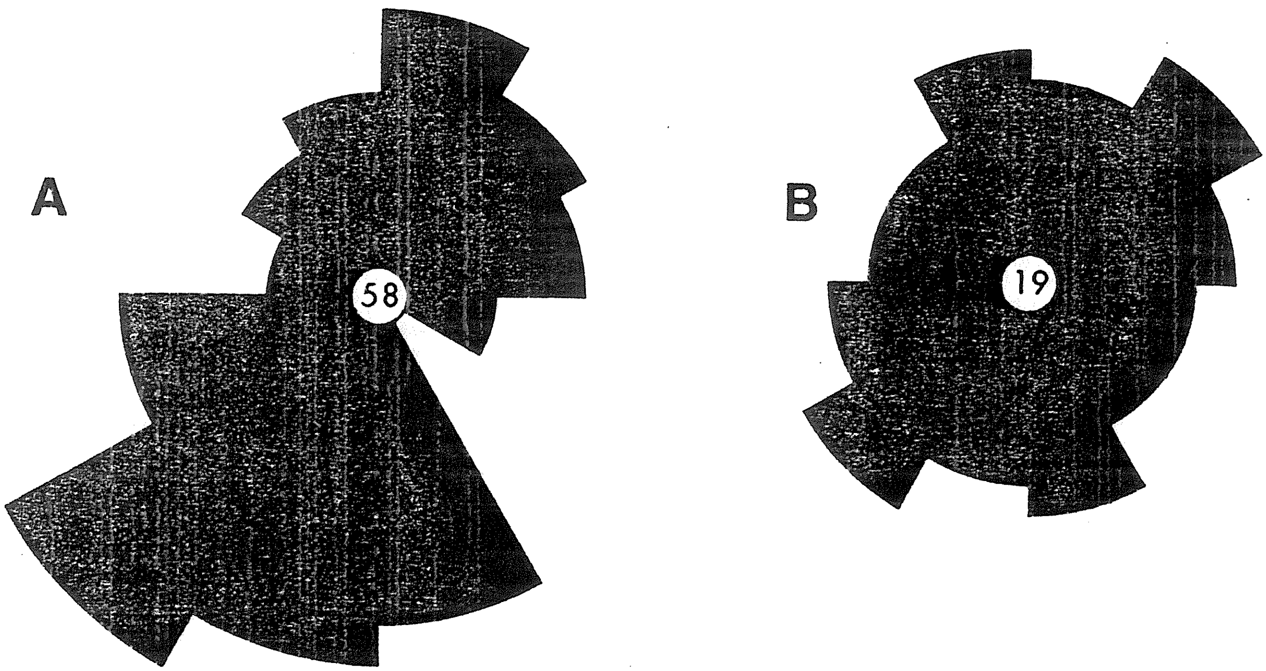


Fig. 13. Equal-area rose diagrams from Grand Bank Head (section A), Facies 1.1 and 1.2: A, ripple paleocurrents (n=58); B, normals to ripple crests (n=19).

Storm processes left a strong imprint on Facies 2.1 and 2.2: (a) thin to medium bedded tempestites record deposition under the influence of combined flows with unidirectional currents dominating the early phase of deposition and combined flow or oscillatory currents dominating the later stages of deposition; (b) flat-pebble conglomerates formed from the erosion and transport of rip-up clasts and rounded shale clasts; (c) lenticular hummocky cross-stratified beds (H-beds) are 'starved' storm-sand units formed under strong swell; (d) gutter casts and pot casts, spectacularly exposed in Facies 2.1, are shallow-water storm features. Pot casts represent the infilling of variably shaped circular to slightly oval depressions formed by the erosion of the substrate by spiralling eddies or whirlpools. Gutter casts are polygenetic features that form by the infilling of elongate scours.

Based in part on the vertical distribution of lithofacies within member 2, a model is proposed for tempestite deposition along fine-grained shorelines (Fig. 14B). This model differs considerably from published models for sandy shorelines (Fig. 14A) in which sand bed thickness is greatest near shore and decreases seaward. In our alternative model (Fig. 14B) the shallow subtidal area consisted of discontinuous, very thin sandstone beds in association with numerous pot and gutter casts -- the Gutter Cast Facies 2.1. This is transitional into deposits of the Siltstone/Sandstone Facies 2.2, characterized by more continuous thin bedded tempestites (Subfacies 2.2A) deposited in a deeper subtidal setting, and medium bedded tempestites and H-beds deposited farther out on the shelf, but above storm-wave base (Subfacies 2.2B). Finally, in the outer-shelf region (typified by member 3) laminae to very thin beds of sand were deposited below the influence of storm waves. The most proximal setting was an area of throughput in which erosion processes predominate. Gutter casts in this part of the shoreline record erosion of the substrate by strong seaward-flowing relaxation currents. In deeper water the throughput/deposition ratio decreased and thicker more extensive sand beds were deposited, including HCS beds which formed under conditions of strong wave swell. Most of the sediment transported out to sea was deposited above storm wave base so that in the outer-shelf area very thin distal tempestites were deposited. Hence, from the shoreline toward the outer shelf, tempestite bed thicknesses first increase, then decrease, with a maximum somewhere between fair-weather wave base and storm wave base.

The abundance and wide variety of marine slides and gravity flows in Facies 2.1 and 2.2 (Table 1) demand unusually high sedimentation rates, necessary to promote triggering of failures on the low slopes typical of the shelf environment. A deltaic setting is strongly suggested by the abundance of these features. The siliceous nature of the fine-grained sediments of the Chapel Island Formation allows slabbing and detailed examination and interpretation of siltstone/mudstone gravity-flow deposits. Slabs will therefore be used as a supplement to field examination at some stops. It appears that support mechanisms in the gravity flows changed both spatially and temporally, and many deposits that will be examined on the trip are intermediate in character between the four end members of the Middleton and Hampton (1973) classification scheme (liquefied flow, turbidity current, debris flow, grain flow).

Member 3 Sedimentology

Member 3 contains only one lithofacies, the Laminated Siltstone Facies 3.1. Sandstone beds in this facies represent distal tempestites deposited by waning flows on the outer shelf below storm wave base. Parallel-laminated sandstone beds with current lineation and parting lineation imply upper-flow-regime conditions. Those beds that lack overlying current ripples are thought to have been deposited from very rapidly decelerating currents. Beds are commonly parallel laminated at their base and ripple cross-laminated above. Data from some of these beds indicate a clockwise shift of paleocurrents during deposition. These beds have been explained as deposits of geostrophic flows in which sediment-laden storm currents are deflected by the Coriolis Effect (Myrow, 1987).

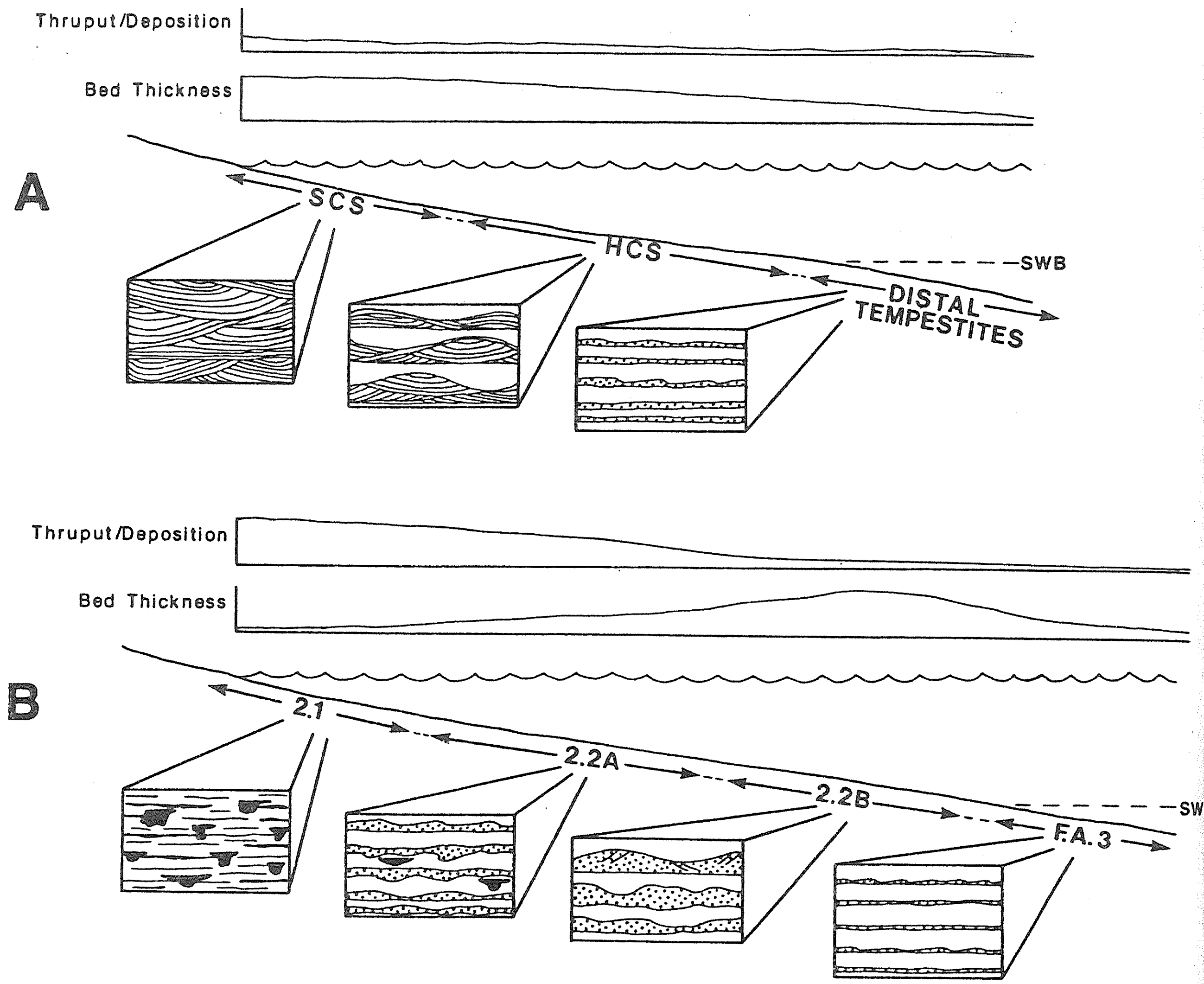


Fig. 14. Comparison of standard tempestite model and the Chapel Island model. Proximality trends in the standard model (A) include constantly decreasing bed thickness away from the shoreline with a shift from SCS to HCS sandstones to distal turbidite-like beds. The Chapel Island model (B) shows a bed-thickness trend that first increases and then decreases away from the shoreline. The proximal setting is one of bypass and significant erosion (gutter casts). Seaward, gutter casts dwindle and bed thickness increases. Rare HCS is formed in the thicker sandstone beds farther out on the shelf. Below storm wave base (SWB), distal tempestites are turbidite-like in character.

BIOSTRATIGRAPHY OF THE CHAPEL ISLAND FORMATION AND THE PRECAMBRIAN-CAMBRIAN BOUNDARY

The discovery of small shelly fossils in the Chapel Island Formation at Dantzic Cove by Greene and Williams (1974) sparked considerable interest in the biostratigraphic relationships exhibited by the Precambrian to Lower Cambrian strata on the Burin Peninsula. The rocks on the Burin Peninsula are considered by the International Precambrian-Cambrian Boundary Working Group as one of three serious candidates for the boundary stratotype (Cowie, 1985), the other sections being located in the Ulakhan-Sulugar region, Siberia, U.S.S.R. and the Meishucun region of southern China. The latter are dominated by carbonate rocks and contain abundant small shelly fossils but relatively impoverished trace fossil assemblages (Cowie and Rozanov, 1983; Xing and Luo, 1984). In contrast, the Burin sections are dominantly siliciclastic, and exhibit the most abundant and diverse assemblage of trace fossils of this time period in the world (Crimes and Anderson, 1985; Narbonne et al., 1987). Small shelly fossils occur sporadically in shale and limestone beds, particularly in member 4 of the Chapel Island Formation (Bengtson and Fletcher, 1983; Narbonne et al., 1987), but are considerably less common and less diverse than in the Chinese and Siberian candidate stratotypes. The Precambrian-Cambrian Boundary Working Group met in St. John's in August, 1987 (Landing et al., 1987) with a post-meeting field trip that examined the upper Precambrian to Cambrian biostratigraphy of the Burin Peninsula and adjacent regions. The result of this meeting was a majority vote by the ten voting members present in favour of the boundary stratotype on the Burin Peninsula. Further deliberations and a postal ballot of the full 31 voting members will follow.

The unique trace fossil succession of the Chapel Island Formation was recognized by Bengtson and Fletcher (1983) and later discussed in detail by Crimes and Anderson (1985). Collaborative studies during the last few years have led to revision of the biostratigraphy of the Chapel Island Formation and formal proposal of a boundary stratotype (Narbonne et al., 1987). Narbonne et al. (1987) and Narbonne and Myrow (in Landing et al., 1987) recognize three globally-correlatable ichnofossil zones in the Chapel Island Formation (Figs 15 & 16). The basal *Harlaniella podolica* Zone (member 1 and base of member 2) contains five genera of simple trace fossils, two of which (*Harlaniella* and *Palaeopascichnus*) occur elsewhere only in Precambrian strata. Shortly above the base of member 2 is the base of the *Phycodes pedum* Zone, which includes the first complex feeding burrows (*Phycodes*, *Gyrolithes*), dwelling burrows (*Skolithos*, *Arenicolites*), anemone burrows (*Conichnus*, *Bergaueria*) and arthropod scratch marks (*Monomorphichnus*). Trace fossils of the *Rusophycus avalonensis* Zone first appear approximately 135 m above the base of member 2 and are common throughout the upper part of the Chapel Island Formation and the overlying Random Formation. The zonal assemblage includes all ichnogenera present in the *Phycodes pedum* Zone plus additional arthropod traces (*Rusophycus*, *Cruziana*, *Dimorphichnus*), dwelling burrows (*Diplocraterion*) and complex feeding burrows (*Taphrhelminthopsis*, *Helminthoida*).

The small shelly fossils of the Chapel Island Formation were first described by Bengtson and Fletcher (1983) and later by Narbonne et al. (1987) and Landing et al. (1987). The organic-walled, worm-like megafossil *Sabellidites cambriensis* first occurs a few metres below the top of member 1 and is found in dark shaly units throughout the lowest 140 m of member 2. The lowest occurrence of small calcareous shelly fossils is in the upper part of member 2, approximately 600 m above the base of the formation, where thin tapering mud-filled tubes of the fossil *Circotheca*? sp. have been recovered (Narbonne et al., 1987). A more diverse *Aldanella attleborensis* assemblage is found in the uppermost part of member 3 and all of member 4. The mudstones in this part of the sequence contain *Circotheca*? sp. and a variety of molluscs preserved as pyritized steinkerns. Peritidal limestones contain a much more diverse assemblage of molluscs, phosphatic problematica and hyoliths that are not present in the mudstones.

Carbonaceous impressions of vendotaenid algae, and impressions of medusoid coelenterates occur sporadically throughout member 1 and the lower half of member 2. G. Vidal (Univ. Lund, Sweden) and M. Moczydowska (Geol. Institute, Poland) sampled the Chapel Island Formation for acritarchs in August, 1987.

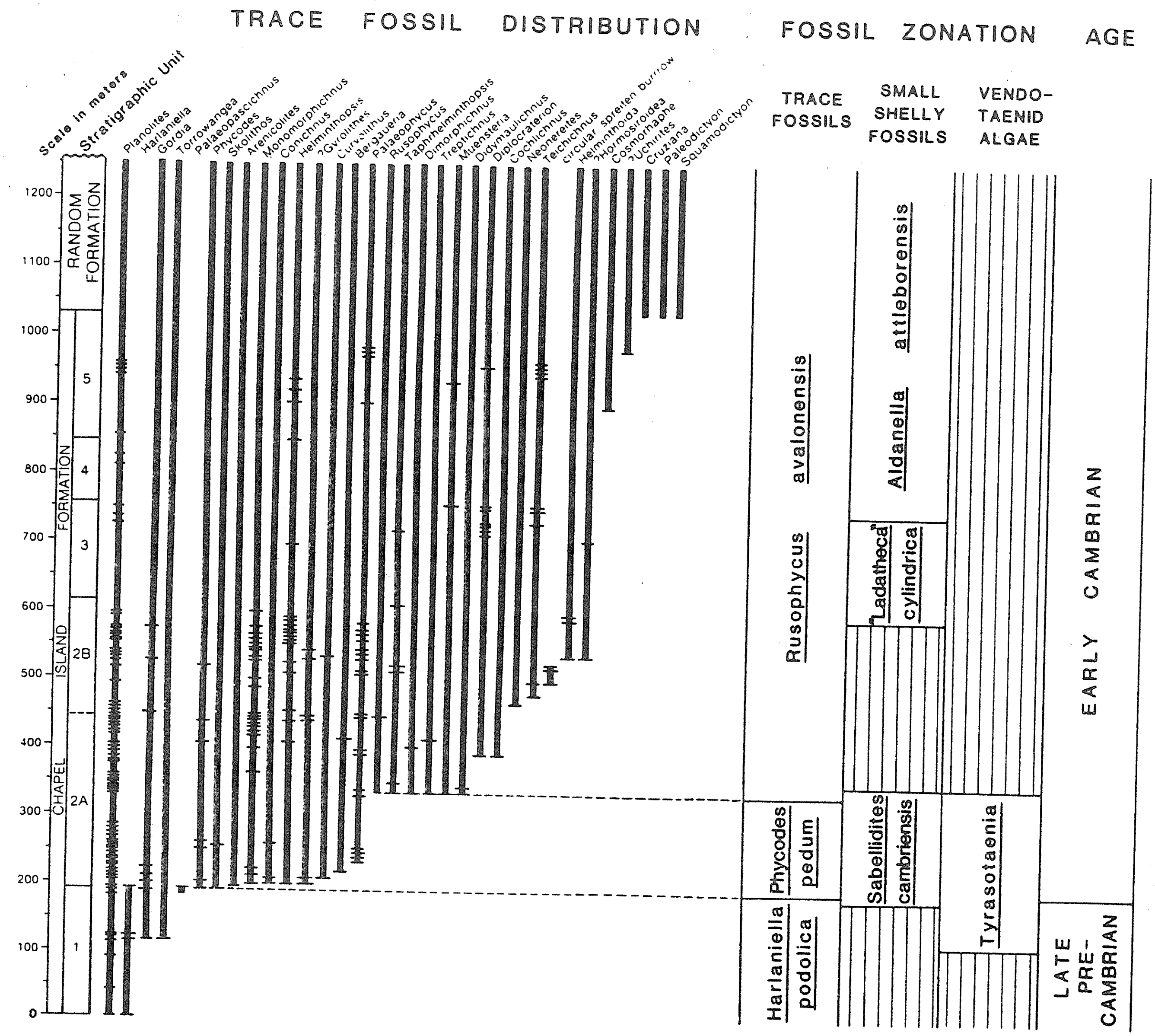
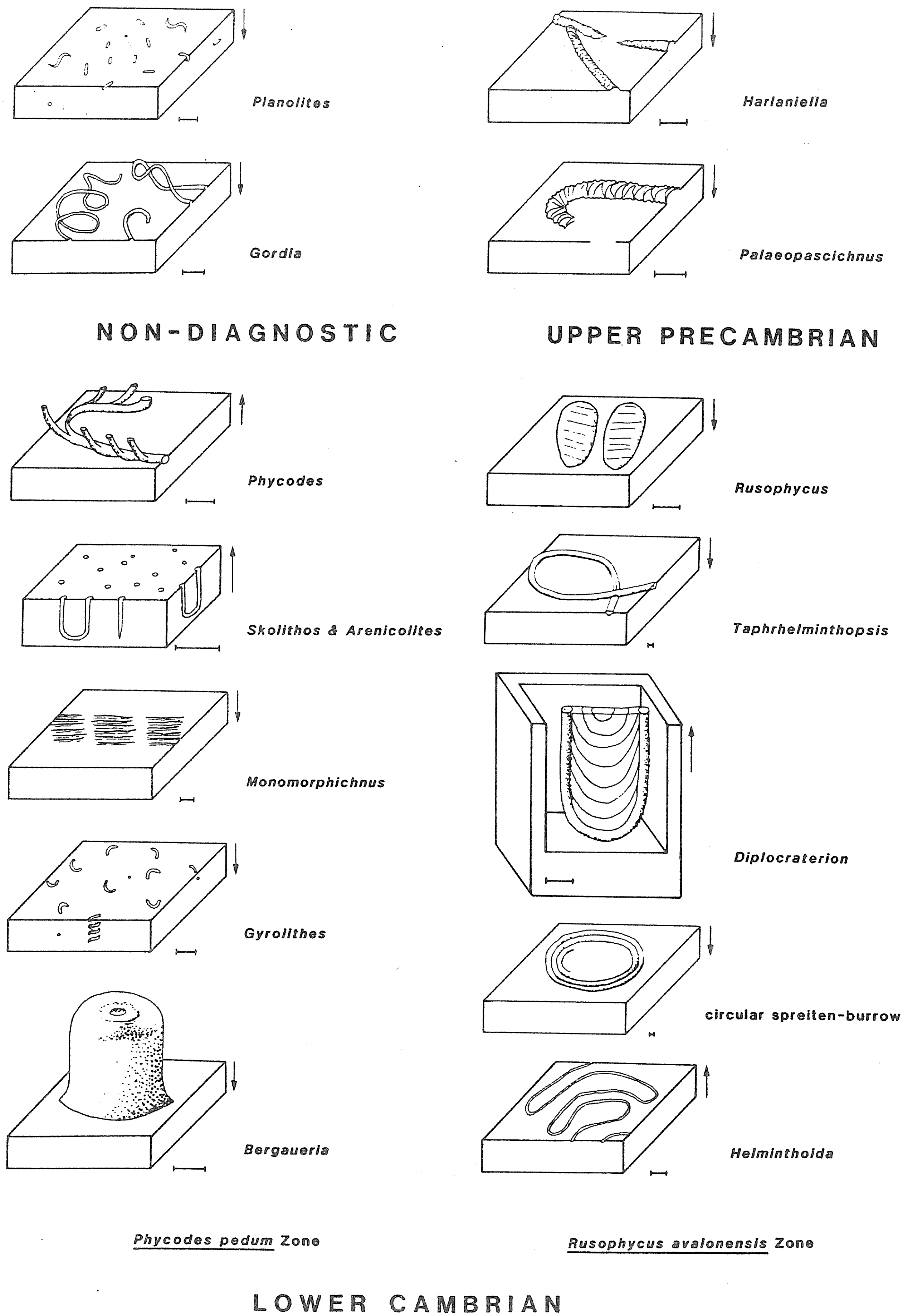


Fig. 15. Fossil zonation and distribution of trace fossils through the Chapel Island and Random formations (after Narbonne et al., 1987, their figure 4).

Fig. 16 (on following page). Typical trace fossils of the Chapel Island Formation. The arrow shows the orientation in the outcrop. Scale bar represents 1 cm.


Planolites
Harlaniella
Gordia
Palaeopascichnus
NON-DIAGNOSTIC
UPPER PRECAMBRIAN
Phycodes
Rusophycus
Skolithos & Arenicolites
Taphrhelminthopsis
Monomorphichnus
Diplocraterion
Gyrolithes
circular spreiten-burrow
Bergaueria
Helminthoida
Phycodes pedum Zone
Rusophycus avalonensis Zone
LOWER CAMBRIAN

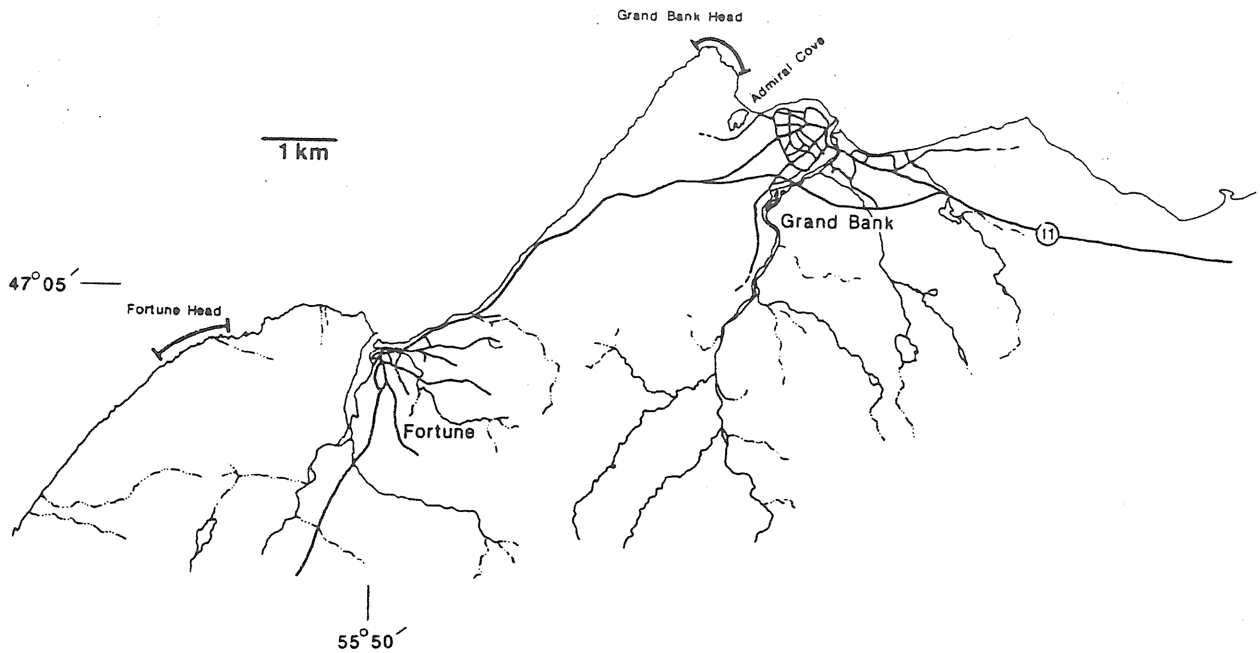


Fig. 17 . Location of Grand Bank Head and Fortune Head sections.

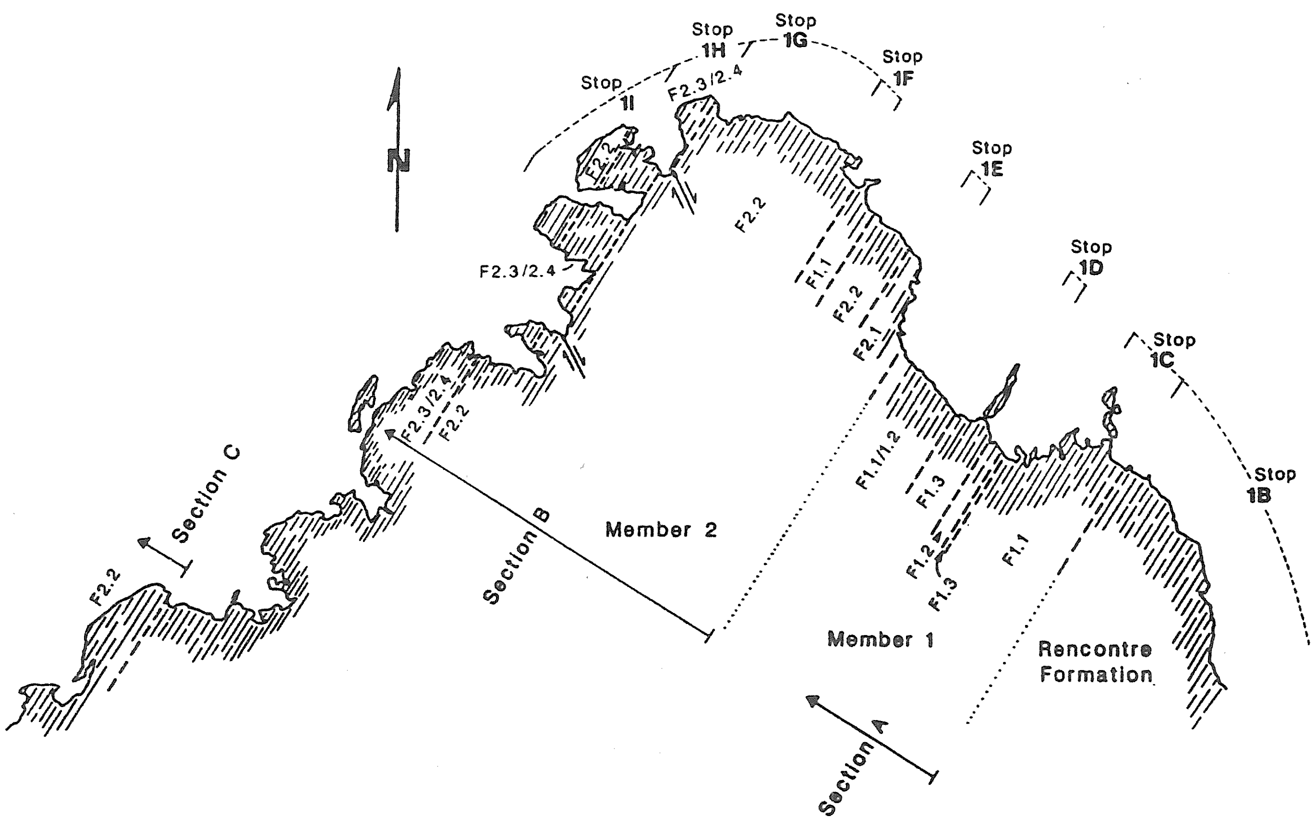


Fig. 18 . Detailed map of the Grand Bank Head section (Stops 1B-1I). Diagonal ruling indicates strike.

GRAND BANK HEAD

Stop 1B

Middle and upper parts of the Rencontre Formation

Description: Over 100 m of strata are exposed along the shore on the west side of Admiral Cove. The outcrop is low lying and easily accessible. The rocks are red, micaceous, fine grained, finely laminated sandstone, siltstones and mudstones. Sedimentary features such as ripple marks, shrinkage cracks, intraformational conglomerate, and ball-and-pillow structures are common. Thin to medium bedded sandstones contain parallel lamination and ripple cross-lamination, as well as scour-and-fill structures and clay-draped scour surfaces.

Interpretation: Depositional environments for the lower part of the formation were probably alluvial/fluvial, but for much of the middle and upper part, particularly near the transition into the Chapel Island Formation, they were more likely marginal marine or peritidal.

Fossils: Despite vigorous searching no fossils have ever been noted.

BACKGROUND FOR FORTUNE DUMP SECTION

An important feature of Facies 2.2 (both subfacies) are disorganized beds; these are very well-exposed at Fortune Dump. The disorganized beds are grouped into four categories, three of which can be seen at this stop.

Disturbed beds are intervals of thinly-bedded strata which have deformed plastically more or less *in situ*, with little evidence for downslope motion. **Unifite beds** are beds of muddy siltstone, 10-100 cm or more thick, that lack obvious internal structure or grading. Polished slabs reveal a range of sedimentary features not visible in outcrop. Beds range from (a) megascopically structureless (Type 1A) to (b) very slightly laminated (Type 1B) to (c) subtly graded and laminated (Type 2), to (d) well-graded and laminated with incipient ripple lamination (Type 3). Figure 19 shows an end-member classification scheme and proposed interpretations for these unifite beds. A continuum of bed style exists between the end members with the interpretation of flow conditions ranging from liquefied flow to fully turbulent flow.

Raft-bearing beds are beds of muddy siltstone, similar in character to Unifite beds, that contain clasts, or 'rafts', of the thinly interbedded lithologies within which they occur. These beds appear to be transitional into both disturbed beds and unifite beds.

Several disorganized beds are found at this stop. At GBB-51.85 a 16 cm-thick bed of coarse siltstone/fine sandstone displays well-developed flow rolls in which the hinge-lines of folds project from the outcrop. An anomalously thick (1.1 m) unifite bed is present at GBB-74.3, and an 80 cm-thick raft-bearing bed is found at GBB-113.8. The latter is interesting in that a 15-20 cm-thick, hummocky cross-stratified sandstone bed directly overlies this bed. The association of a mass-movement deposit with storm-generated structures is also found at FD-346.2 (Stop 2M), and implicates storm-generated, cyclic wave-loading as a possible contributor to slope failures in Facies 2.2.

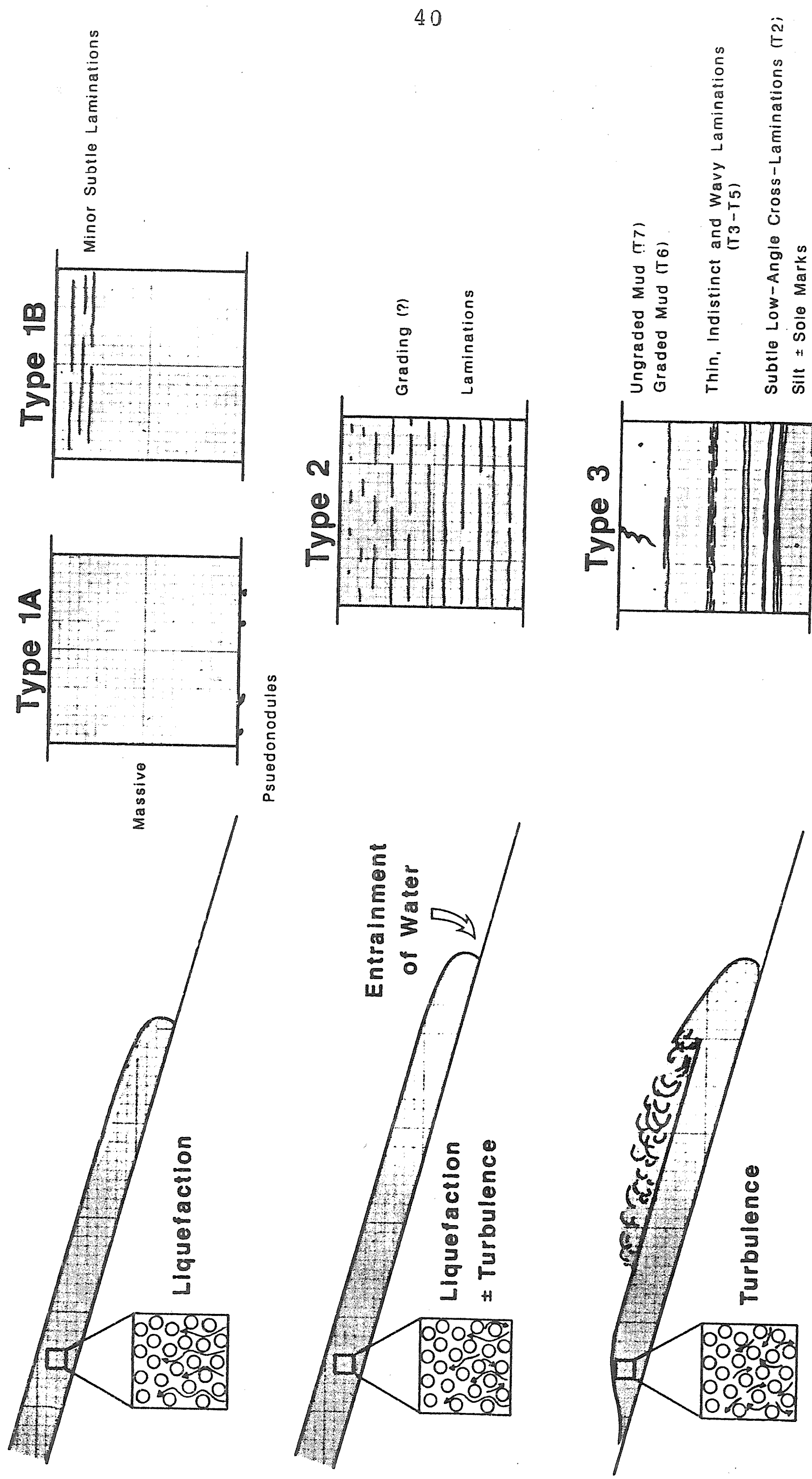


Fig. 19 . Unifite classification scheme. Characteristic features are at right. Turbidite divisions of Type 3 Unifites are those of Stow and Shanmugam (1980).

Interpretation: Evidence indicates that the bulk of Facies 2.2 was deposited in nearshore and inner-shelf settings (between the shoreline and storm wave base). This is based on the close association with facies deposited at the shoreline (with evidence of periodic exposure), the presence of HCS, and the abundance of wave ripples and wave-ripple laminae. The upward transition in sandstone beds from parallel laminae to combined-flow or wave-generated laminae is a fundamental distinguishing feature of tempestite beds (e.g., De Raaf et al., 1977; Goldring and Bridges, 1973; Kreisa, 1981; Aigner, 1982; Seilacher, 1982; Brenchley, 1985).

Figure 22 gives an interpretation for the origin of the disorganized beds, including proposed flow transitions for the downslope evolution of one bed type to another. Many beds have characteristics of debris flows, namely matrix support, apparently random fabrics, variable clast size, rip-up clasts, and internal flow structures (Nardin et al., 1979).

The abundance of nearshore instability features (i.e., disorganized beds) requires high sedimentation rate to generate conditions suitable for failure (high pore pressures favourable for liquefaction) on the slopes typical of nearshore marine settings. According to Moore (1961, p. 351), failure of metastable (underconsolidated) sediments is "... normally found in the marine environment only off large rivers carrying predominantly fine-grained silt and clay-sized material". Facies 2.2 is considered to be an inner (Subfacies 2.2A) and intermediate (Subfacies 2.2B) shelf deposit formed in an area receiving large quantities of fine-grained sediment, and one subjected to repeated storm events.

Fossils: Trace fossils are abundant and include Cambrian-type forms of *Phycodes*, *Monomorphichnus*, *Treptichnus*, *Diplocraterion*, *Helmenthopsis*, and very well preserved *Rusophycus*. The strata can be assigned to the *Rusophycus avalonensis* Zone.

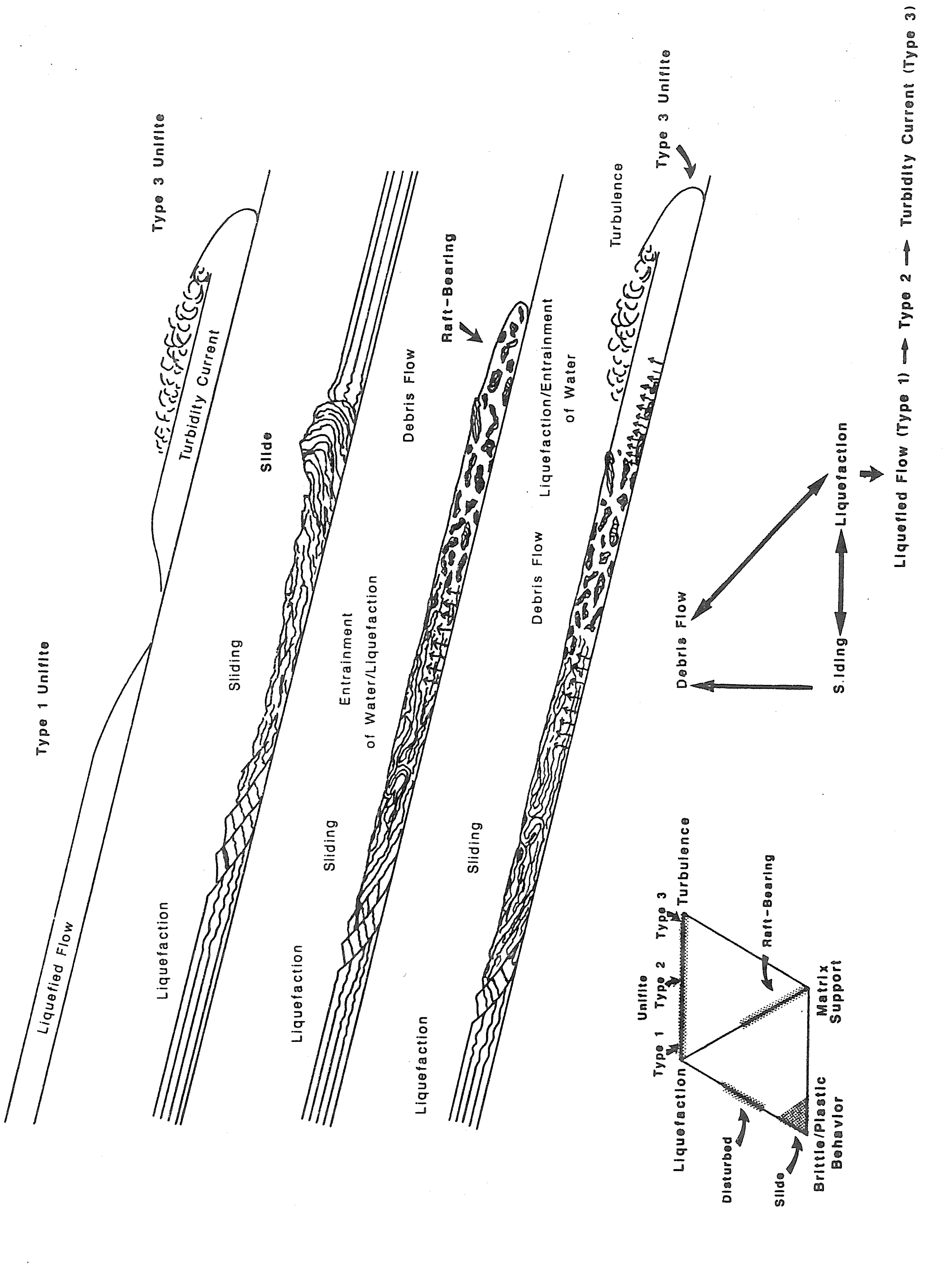


Fig. 20. Hypothetical flow transformations for disorganized beds, depicted graphically and summarized (lower right). Quadrilateral in lower left shows interpreted support mechanisms for the different bed types at the time of deposition.

FORTUNE HEAD

Fortune Head lighthouse is reached by a gravel road 2.7 km south of the town centre on Route 220. Park vehicles at the fork in the road near the lighthouse. The base of the Fortune Dump section lies several hundred meters east of the lighthouse. The section extends west along the coast for approximately 1 km (Fig. 21). The 400 m-thick section (Fig. 22) is superbly exposed in a continuous series of low cliffs that are readily accessible by foot. The strata dip northwest at 15-46° with dip increasing up-section. A few faults are present, but marker horizons allow easy correlation across them.

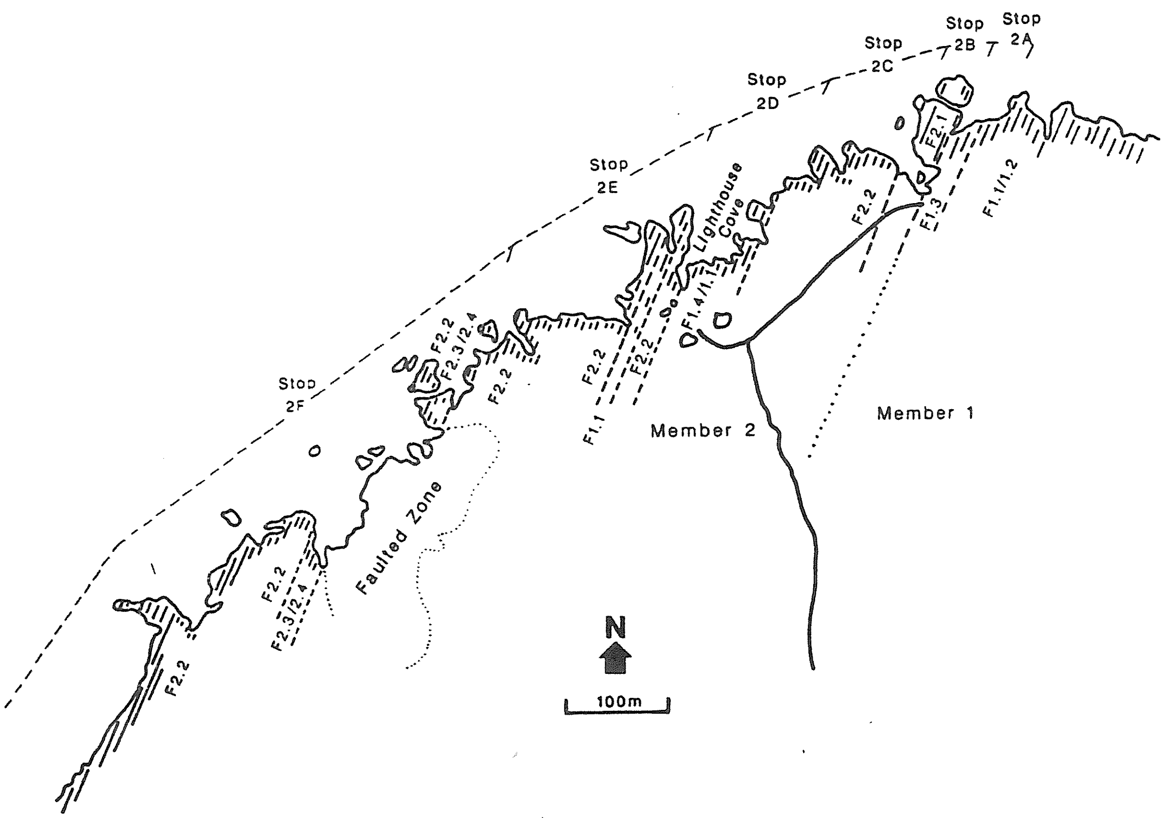


Fig. 21 . Map of Fortune Head. The Town Dump is in the cove at Stop 2C. Diagonal ruling indicates strike. Stops 2A-2F only are shown.

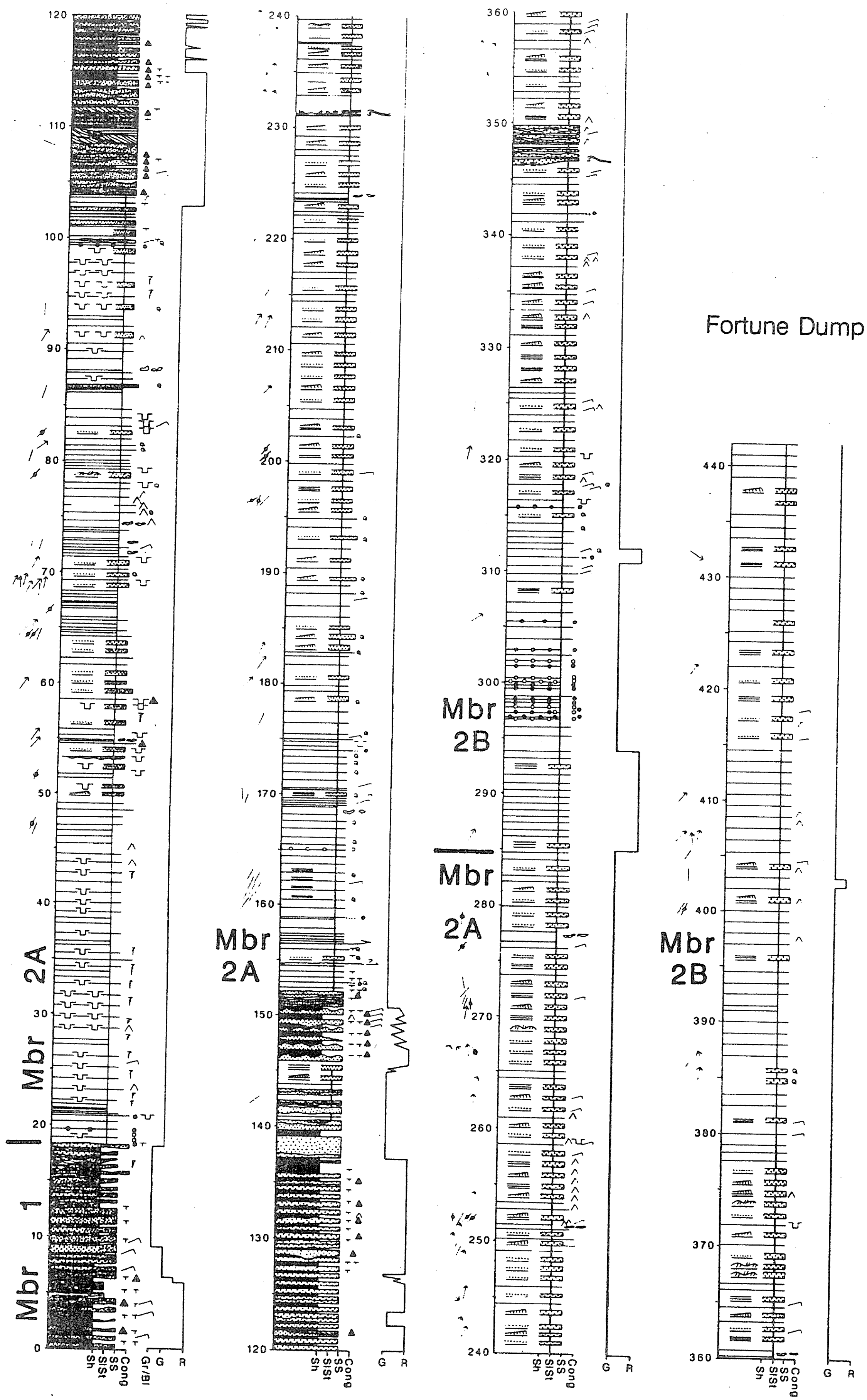


Fig. 22 . Schematic section at Fortune Dump. Key is in Figure 15.

Stop 2A

Upper part of member 1 of the Chapel Island Formation

Descend a small gully on the east side of a promontory located just east of a small cove that acts as the Town Dump. Shallowly-dipping redbeds of Facies 1.1/1.2 may be difficult to reach at high tide. Black shales and gray silty shales of Facies 1.3 form the uppermost part of member 1. These are cut by a minor fault in the notch of the cove. The top of member 1 is located on the prominent bench of the promontory. **Strata are slippery when wet.**

Description: Many of the features of Facies 1.1/1.2 are readily observed at the base of the section, including synaeresis and desiccation cracks and abundant current ripples. The rocks of Facies 1.3 are poorly weathered below the small fault, but contain abundant synaeresis and a few desiccation cracks. The member 1-2 boundary has a number of interesting features including desiccation cracks and abundant phosphatic shale clasts; at the end of the promontory there is a shallow elongate scour that is covered with phosphatic shale clasts up to 15 cm in length. Ripples on the bedding surface adjacent to the scour continue down into the scour, and based on their tuning fork junctions are considered wave ripples.

Interpretation: See the interpretations of these facies at Stops 1C and 1D. The prominent bench is considered to be a minor disconformity surface that shows evidence of subaerial exposure and reworking by waves after submergence.

Fossils: Fossils are present sporadically in these beds, mostly in the gray-black units. Trace fossils are simple forms such as Planolites and Gordia. Well-preserved examples of the carbonaceous small shelly fossil Sabellidites cambriensis and the vendotaenid alga Tyrasotaenia occur in the darker, shalier units. One impression of a soft-bodied ("medusoid") organism was found near the top of these strata. The overall composition of the fauna and flora suggest a latest Precambrian age.

Stop 2B

Lowermost member 2 and the Candidate Precambrian - Cambrian Boundary Stratotype Point (FD-18.6-35)

The basal part of member 2 can be viewed along the prominent bench, but the rest of the section must be viewed from above. **The candidate Precambrian-Cambrian boundary stratotype horizon and point is located 2.4 m above the base of member 2.**

Description: The strata belong to the Gutter Cast Facies 2.1; i.e., mainly laminae to very thin beds of green-gray sandstone and silver-green siltstone, with moderate to abundant gutter casts. The term **gutter cast** refers to erosional structures that range from downward-bulging sole structures to isolated channels. They are composed of fine sandstone; less commonly of white-weathering, well sorted, fine to medium grained sandstone and granule to pebble conglomerate. The sandstone gutter casts are either massive, planar-laminated, or contain oscillatory-flow laminae. The thicknesses of the gutter casts are generally one or more orders of magnitude greater than the thickness of the sandstone beds with which they are intercalated. Carbonate nodules are found within these beds, and in many cases the entire bed experienced early carbonate cementation.The sandstone content of this facies, excluding gutter casts, is 10-40%. Quartzitic beds make up about 5% of this facies. Sandstone beds greater than a few centimeters are rare. Unifite beds (Types 1 and 2; Figure 21) are prominent, forming 10-15% of the section, and one or two raft-bearing beds have been noted. Ptygmatically folded sandstone dikes are surprisingly abundant, in some cases leading to gutter casts. Wave ripples are present on thin sandstone beds, locally as starved forms.

A wide variety of gutter-cast shapes are seen both in cross-section and in plan view. Extremely steep to overhanging walls are common. Compaction has altered the shape and internal stratification in some of these beds. Sole markings on the sides and bases of these beds include groove marks, poorly developed prod and flute marks, and post-depositional trace fossils. Groove marks are generally parallel to subparallel to the long axes of the gutter casts and circular to downward-spiralling on pot casts.

The long axes of gutter casts at FD show a strong northeast-southwest orientation. Some gutter casts are quite sinuous. Numerous examples of bifurcating gutter casts were noted, as well as those that taper and pinch out along strike. All of the bifurcating examples have their "forks" opening toward the northeast, and nearly all pinchouts are toward the northeast as well. The upper bedding surfaces of some gutter casts are covered with symmetrical ripple marks. The crests of these ripples are generally perpendicular to the trend of the gutter cast, but beds with multiple directions occur.

In a few cases, gutter casts are seen originating from, and leading into, **pot casts**. These pot casts commonly weather out as partially free-standing structures. They range from discs to rounded loaf-like forms to tall pillars and from remarkably small (1 cm diameter) to very large (ca. 20 cm diameter). They commonly widened downward, and a few examples have a snail-like or corkscrew shape similar to potholes found in bedrock along modern rivers. The bottoms of pot casts are commonly deepest around the outside, with a central erosional high: the form resembles the base of a wine bottle.

Unifite beds are graded to nongraded siltstone and silty mudstone characterized by a lack of obvious internal structure in the field. Cut and polished slabs, however, reveal a range of subtle internal structures (Fig. 19). The bases of these beds are very sharp, and the tops vary from gradational to very sharp. These beds range from 10-70 cm thick, averaging 36 cm at this exposure. These beds are normally tabular, but locally have a channelled margin (e.g. FD-27.3).

Interpretation: Intimate association with peritidal redbeds and the shoreline deposits of Facies 1.3 indicates a shallow subtidal environment. The gravity flows suggest high sedimentation rates and deposition in deltaic environments. High sediment accumulation rate is supported by the abundance of sedimentary dikes.

Data from this outcrop prompted the model for tempestite deposition along fine-grained shorelines (Fig. 14). The Gutter Cast setting represents a zone of bypass, an area dominated by erosion and throughput of sediment. Perhaps fast-moving sediment flows were debouched from distributary channels into the shallow subtidal environment, where they eroded the sea floor on entry into the marine setting and then deposited their load farther out across the delta-front, prodelta and inner shelf.

Observations from the gutter casts in this facies, including steep to overhanging side walls, suggest a short period of time between cutting and filling (see Goldring and Aigner, 1982). Overhanging walls must have been very common in potholes because most pot casts are tilted and/or widen downward. These potholes were eroded into a semi-cohesive to cohesive substrate of clayey silt by stationary eddies carrying sand in a manner similar to that outlined for bedrock potholes (Alexander, 1932). The thickness of the gutter casts (generally 10-100 times thicker than the associated sandstone beds) and the evidence for rapid infilling of the gutters is compatible with deposition from thick sediment-laden flows that deposited sediment almost exclusively in the gutters. The wave ripples on the top of several gutter casts formed during the late stages of deposition or from later reworking prior to burial.

Unifite beds (Fig. 19) have grain sizes (dominantly silt) particularly susceptible to liquefaction and downslope flow. The source may have been muddy delta-front mouth bars. The well graded and stratified unifites (Type 3, Fig. 19) may have been generated closer to shore, allowing evolution into a turbidity current prior to deposition on the delta front. Excellent analogues for these silty flows are described by Prior et al. (1986) from the Huanghe (Yellow River) Delta. They describe silt-flow gullies that are filled with acoustically transparent sediments thought to have been generated by liquefaction.

Fossils: Fossils are abundant and can be used to define a **Precambrian-Cambrian boundary** (Fig. 23). Member 1 and the basal 2.4 m of member 2 represent the uppermost strata of the *Harlaniella podolica* Zone. This zone is characterized by a low diversity trace fossil assemblage dominated by simple subhorizontal burrows (*Planolites*, *Gordia*, *Harlaniella* and *Palaeopascichnus*). The latter two ichnogenera are particularly valuable age indicators as they apparently are restricted to the upper Precambrian. The vendotaenid alga *Tyrasotaenia* is also common in these strata.

Phycodes pedum first appears 2.4 m above the base of member 2, and can be used to define the base of the *Phycodes pedum* Zone. This point has also been proposed as a candidate stratotype point and horizon for the Precambrian-Cambrian boundary (Narbonne et al. 1987). The proposed boundary horizon occurs within a continuous and relatively uniform succession of subtidal strata. The highest occurrence of Vendian ichnofossils (*Harlaniella podolica* and *Palaeopascichnus delicatus*) is only 0.2 m lower in the section, in the same lithofacies.

Higher strata of member 2 at this stop contain elements of the *Phycodes pedum* Zone including arthropod traces (*Monomorphichnus*), vertical dwelling burrows (*Skolithes* and *Arenicolites*) and coelenterate resting burrows (*Conichnus*). Vendotaenid algae also occur sporadically in these strata. In general, the faunal assemblage is closely similar to that present in the beds of the *Phycodes pedum* Zone at Grand Bank Head (Stop 1E).

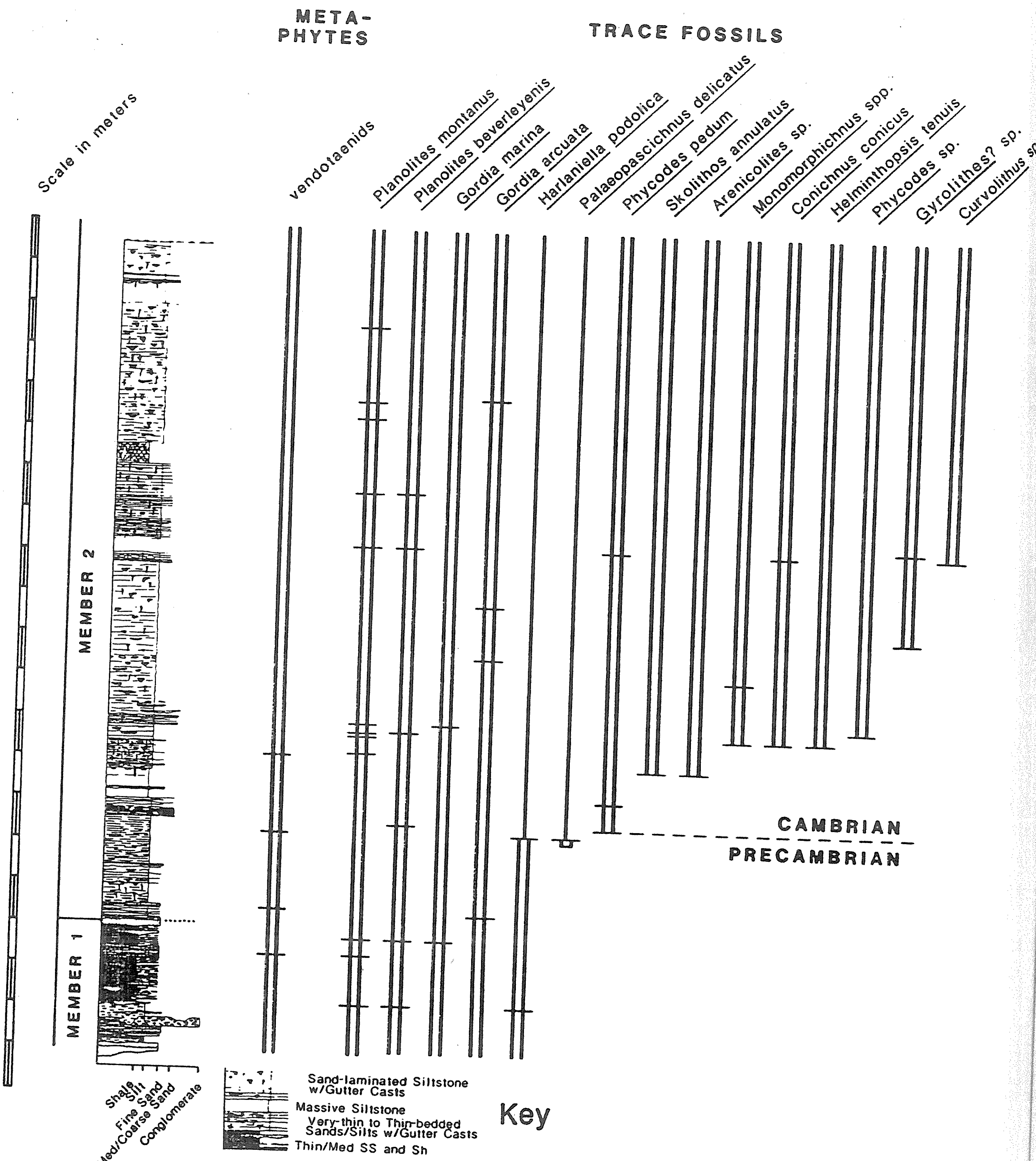


Fig. 23. Distribution of fossils across the candidate Precambrian-Cambrian boundary stratotype horizon at Fortune Head (after Narbonne et al., 1987, their figure 5).

DAY 2

GEOLOGY OF THE NORTHERN BURIN PENINSULA: GRANITES, VOLCANIC ROCKS AND ALTERATION ZONES

PREAMBLE:

On the return trip to St. John's, we will briefly examine several aspects of the magmatic history and styles of mineralization characteristic of the Avalon Zone. Stops 2-1, 2-2 and 2-5 will examine the lithologies and gold mineralization in the Love Cove Group, the major late Precambrian Avalonian volcanic succession west of the Paradise Sound Fault. The Hickey's Pond belt will be observed where it crosses the Monkstown Road. This belt is an 80-km-long zone of high-alumina and silica alteration with significant gold values in places (up to 6 g/t). At the Monkstown Road showing, a value of 8 g/t has been reported but this has not been duplicated by resampling (Tuach, 1987). Stop 2-3 and 2-4 will briefly examine examples of Paleozoic and Precambrian granites, respectively. The post-tectonic Devonian Ackley Granite (Stop 2-3) intrudes across the regional fabrics of the Love Cove Group, and contains tin- and tungsten-bearing greisens 25 km west of here. The pretectonic Swift Current Granite (Stop 2-4) is an example of the late Precambrian granites that are characteristic of the Avalonian terranes. The alteration and related gold mineralization in the Love Cove Group is interpreted to be related to the intrusion of this pluton.

If time allows, we will briefly examine aspects of the volcanology and mineralization with some of the youngest Precambrian volcanic rocks, namely, the Bull Arm Formation of the Musgravetown Group, exposed on the Isthmus of Avalon.

EPITHERMAL ALTERATION AND MINERALIZATION OF THE HICKEY'S POND BELT

Areas of high-alumina, argillic, advanced argillic and silicic alteration occur in many parts of the Avalon Zone (Figure 24). Many of these areas were known to occur since the early days of geological work in the area. However, the relationship between these alteration zones and precious metals, although previously mentioned (e.g. Hussey, 1978; O'Brien and Taylor, 1979), was demonstrated only within the last few years (e.g. McKenzie, 1983; O'Driscoll, 1984; O'Driscoll and Huard, 1984; Huard and O'Driscoll, 1985, 1986; Sparkes, 1986 and O'Brien and Knight, 1987, 1988). It is generally thought now that these alteration zones are part of epithermal - fumarolic systems. Epithermal models (Figure 25) which were developed in the western and eastern United States (e.g. Spence et al., 1980; Buchanan, 1981; Berger and Eimon, 1983; Schmidt, 1983, 1985) can be applied to the Newfoundland Avalon Zone and the same signatures can be recognized.

New areas of widespread silicic alteration, which contain significant gold values, have been delineated (Tuach, 1984; Huard and O'Driscoll, 1985, 1986; Sparkes, 1986; O'Brien, 1987; O'Brien and Knight, 1987, 1988). Exploration in these areas is still in its infancy.

The Hickey's Pond gold prospect is located on the Burin Peninsula in an 80-km-long zone of high-alumina and silica alteration that extends from Hickey's Pond to the Knee (Figure 24). The zone contains pyrophyllite, alunite, sericite, specularite, lazulite, chloritoid, apatite, rutile and pyrite (Huard

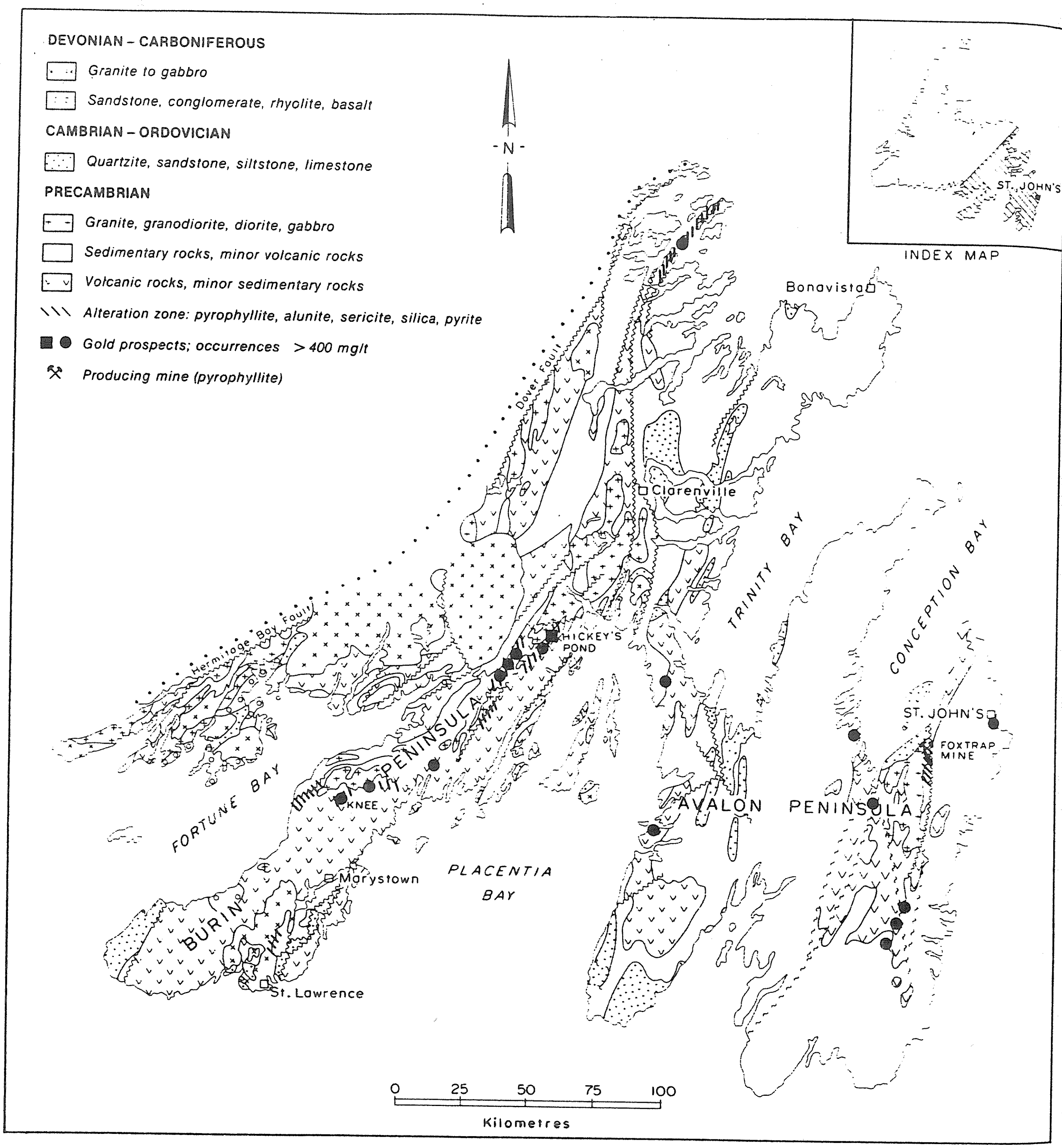


Figure24. Simplified geology of the Avalon Zone in Newfoundland with known gold prospects, occurrences and major alteration zones.

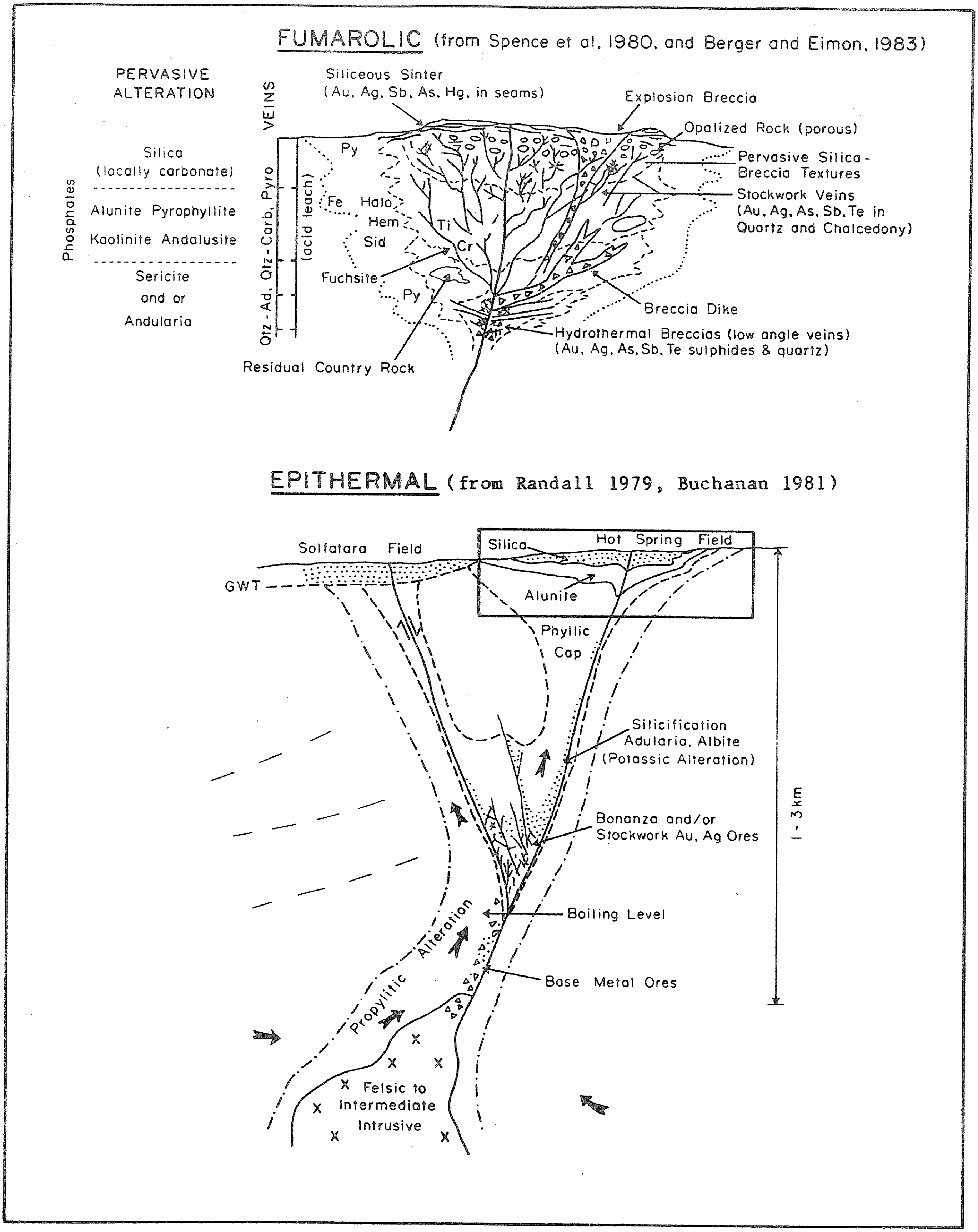


Figure 25. Idealized alteration and structural features in epithermal-fumarolic systems.

and O'Driscoll, 1985, 1986). The prospect is hosted by banded and hydrothermally brecciated hematite- and alunite-bearing rocks in a zone of advanced argillic alteration in Hadrynian volcanic rocks of the Love Cove Group. The deposit is separated from the Late Hadrynian Swift Current Granite by a fault. The alteration zone and the surrounding rocks exhibit a penetrative tectonic fabric and a low-greenschist-facies metamorphic assemblage. Gold values up to 6 g/t have been obtained from specularite-rich breccias in the Hickey's Pond prospect.

The alteration assemblages and depositional features at Hickey's Pond led Huard and O'Driscoll (1986) to conclude that the deposit formed in an epithermal-fumarolic environment. Deposits throughout the belt vary from sinters and chemical precipitates with geyserite eggs to deeper-level hydrothermal stockworks and breccias. Pervasively altered volcanic rocks are ubiquitous.

Minor gold occurrences are found along the belt at Monkstown Road and at the Knee. The Monkstown Road showing will be visited during this field trip. Other areas of silica-sericite-pyrite alteration and anomalous gold content (greater than 0.4 g/t Au) have also been reported from other parts of the Avalon Zone (Figure 5; Sparkes, 1986; Tuach, 1987; O'Brien, 1987; O'Brien and Knight, 1987, 1988). A high-alumina alteration zone just north of St. Lawrence (Figure 5) on the Burin Peninsula consists of pyrophyllite, dumortierite and andalusite. Very little detailed work has been done on this deposit and it is not known if it contains associated silicification and brecciation or anomalous gold values.

For further discussion of the deposits in the Hickey's Pond Belt, see Huard and O'Driscoll, 1986 (handout).

DAY 2: FIELD TRIP STOPS

Stop 2-1: Monkstown Road Showing (waterproof boots are recommended)

The Hickey's Pond Belt (Figure 26) is an 80-km-long zone of high-alumina and silica alteration that extends from Hickey's Pond to the "Knee" of the Burin Peninsula (Figure 24). Gold values up to 6 g/t have been obtained in hydrothermal breccias at the Hickey's Pond prospect. Anomalous gold values (greater than 0.4 g/t) have been recorded in various parts of the belt (e.g. Huard and O'Driscoll, 1985, 1986; Tuach et al., 1988).

The Monkstown Road showing is located where the Monkstown Road crosses the Hickey's Pond Belt. The showing consists of a central area of silicified rocks with an envelope of sericitic and pyritic alteration. At the roadcut we can see intensely sericitized volcanic rocks with thin pyritic cherty bands or aphanitic dykelets. A grab sample from one of these yielded a gold value of 8 g/t. However, this has not been duplicated by resampling (Tuach, 1987).

The main silicified zone can be seen in a swamp approximately 100 m north of the road. It is an extremely siliceous, north-trending elongated outcrop approximately 40 m by 10 m. Most of the outcrop is tan-colored, fragmental, siliceous rock. Fragments are angular and range up to 10 cm across. The matrix is brown quartz, which commonly displays calcedonic banding. These bands persist for only short distances, on the order of 10 cm, and may represent either fragmented fissure fillings in the subsurface or siliceous sinter at

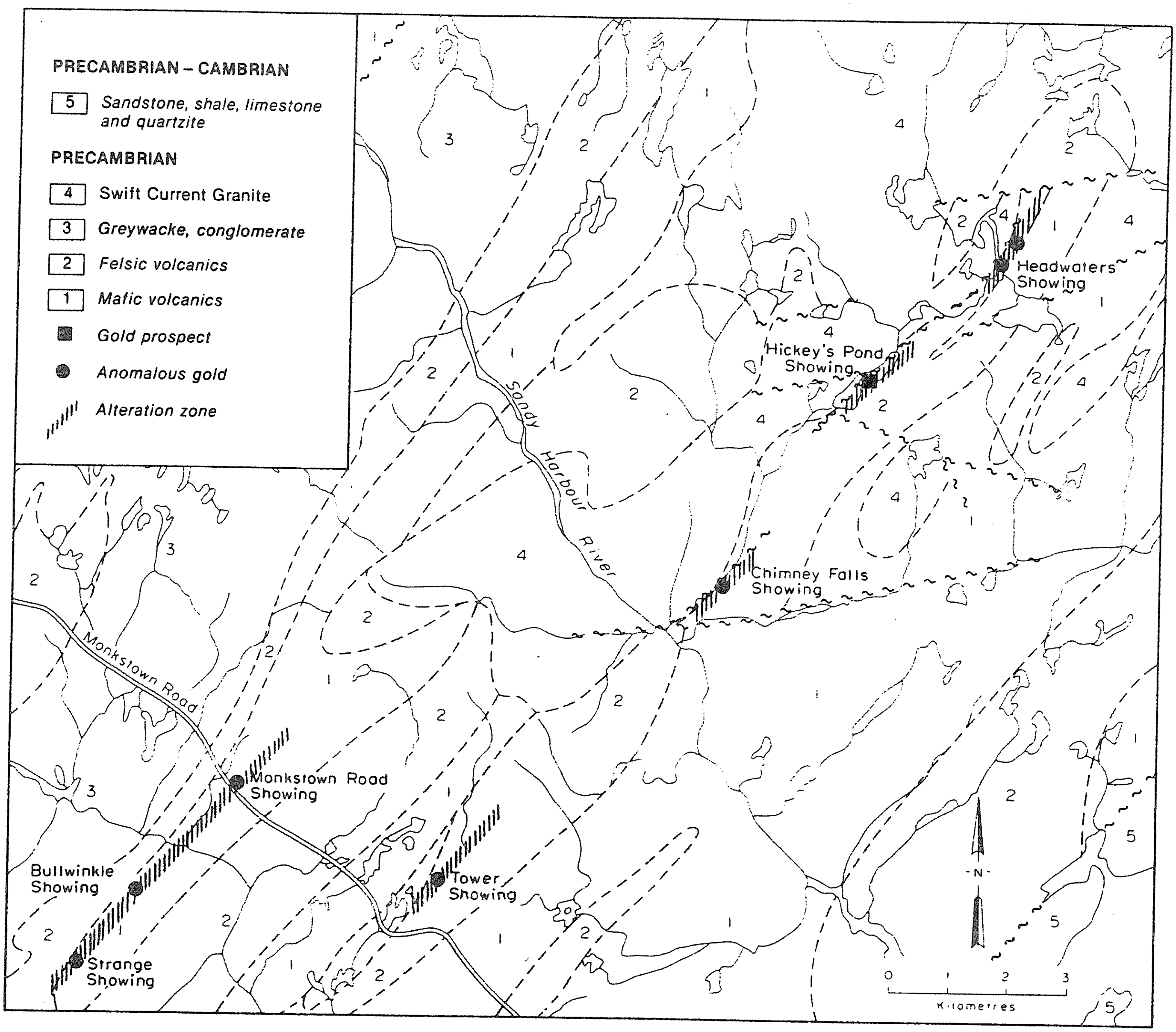


Figure 26. General geology of the Hickey's Pond Belt showing major alteration zones and gold occurrences.

the surface. A small zone of hydrothermal breccia with a specularite-rich matrix is evident on the southwest corner of the outcrop. The brecciation passes upward into thin veinlets of specularite. Specularite and pyrophyllite are also present in coarsely crystalline quartz veins and segregations. Brilliant-blue lazulite occurs in similar quartz-specularite veins throughout the showing. Although alunite is common along the belt and occurs in outcrops about one kilometre to the north and south, it has not been identified at the Monkstown Road showing. Other minerals, which occur throughout the belt, include chloritoid, apatite and rutile.

Huard and O'Driscoll (1986) concluded the rocks in the Hickey's Pond Belt represent various parts of the highest level of an epithermal system (cf. Berger and Eimon, 1983). Rocks in some areas may represent deposits at the surface with chemical precipitates, breccia aprons and geyserite eggs. These are underlain by rocks which have undergone widespread alunitization, silicification, specularitization and pyrophyllitization. Deeper levels are represented by hydrothermal stockwork breccias in altered volcanic rocks.

Stop 2-2: Clastic facies of the Love Cove Group

In the north-central Burin Peninsula, several major clastic intercalations occur towards the top of the basal volcanic succession of the Love Cove Group. These have a varied lithology and include immature arenites with interbedded granule to pebble sandstones and conglomerates.

The dominant rock types at this locality are poorly sorted pebbly sandstones and conglomerates. Planar and cross-bedding are characteristic features.

Stop 2-3: Ackley Granite

The Ackley Granite (cf. Dickson, 1983) is a large irregular shaped pluton, some 3600 km2 in area, which is intrusive post-tectonically into late Proterozoic volcanics and sediments of the Avalon Zone and pre-Middle Ordovician metasediments and Devonian and older granitoids of the Gander Zone. The Ackley Granite consists primarily of medium to coarse grained, uniform to porphyritic, massive biotite granite. In detail, the Ackley Granite is composed of three texturally distinctive varieties of granite. The Ackley Granite truncates the Dover-Hermitage Bay Fault and is dated isotopically (40Ar/39Ar hornblende, Dallmeyer et al., 1981) at 355 ± 10 Ma. Rb-Sr and K-Ar ages concur with this date.

The exposure to be examined at this stop is coarse-grained porphyritic biotite granite which is typical of approximately 75% of the batholith. At this locality the granite is fresh and massive with well developed perthitic potash feldspar phenocrysts. Sphene and rare hornblende also occurs in this outcrop. Locally, small mafic xenoliths and quartz-feldspar pegmatites may be observed.

At this stop also, the contact between the posttectonic Ackley Granite and metasediments of the Love Cove Group is exposed (north end of the roadcut on the western side of the road; southern end of the outcrop on the east side of the road). The sediments have been metamorphosed within an aureole less than 200 m wide with the development of biotite, and locally muscovite, cordierite and andalusite porphyroblasts (not all visible at this outcrop).

Massive sandstones have been recrystallized to produce a hard hornfels whereas recrystallization of laminated sandstones has produced a gneissic banding. The sandstones exposed at this outcrop also contain isolated granitic veinlets which are probable the result of local anatexis (Dickson, 1983). At the soutern end of the outcrop, medium grained porphyritic biotite granite is in contact with the sediments. A few metres into the batholith the granite becomes coarser grained with well-developed rapakivi potash feldspar phenocrysts.

Skyline ridges seen to the west are roof pendants in the Ackley Granite. They consist of volcanic and overlying shallow marine sedimentary rocks of the Long Harbour Group.

Stop 2-4: Swift Current Granite

The Swift Current Granite is a composite pluton with a compositional range from biotite-hornblende granite through granodiorite to syenite with minor aplite. Diorite and granodiorite occur as its marginal phases. It is intrusive, almost exclusively, into the Love Cove Group and bears textural similarities to volcanics adjacent to its margins. The granite is interpreted to be comagmatic with the adjacent volcanics. The Swift Current Granite is pretectonic and hosts a locally strong schistosity. The pluton has been metamorphosed under greenschist facies conditions during Acadian orogenesis (Dallmeyer et al. (1983). The Swift Current Granite has been dated isotopically by several methods and results are given by Dallmeyer et al., 1981. A 580 ± 20 Ma U-Pb zircon date on the pluton gives the approximate time of crystallization. 40Ar/39Ar (plateau) ages of 560-566 Ma on hornblende from the granite are interpreted as dating times of post-magmatic cooling. Within analytical accuracy, U/Pb and 40Ar/39Ar ages overlap, suggesting rapid post-magmatic cooling and shallow intrusive levels.

Two exposures will be examined at STOP 2-4. At the first exposures, near the village of Swift Current, the Swift Current Pluton on the north side of the road is composed mainly of fine to medium grained diorite and gabbro, intruded by medium grained hornblende-biotite granite. Well preserved magma mixing features can be seen in these outcrops. In places, the granite is coarser grained near the contacts with the mafic phases where large (8 cm) acicular hornblende crystals are developed. A similar grain-size coarsening of quartz and potash feldspar can be seen. A possible interpretation of these features is that the mafic phases were dehydrated during the intrusion of later granitic phases, allowing easy contamination of the latter by mafic mineral elements. Slow cooling of both phases permitted growth of large crystals in the granite. In places a faint banding can be seen which may be the result of crystal settling.

At the second exposure, the pluton consists of medium grained hornblende-biotite granite cut by diabase dikes. The granite is metamorphosed and deformed; the intensity of the foliation varies from mild to intense with very localized development of mylonitic zones. Diabase dikes, aligned parallel to cleavage, are completely altered to chlorite schists.

Stop 2-5: Love Cove Group

The name "Love Cove Group" was introduced by Jenness (1957) to designate the metavolcanic rocks which underlie the Late Proterozoic sediments of the

Connecting Point Group in Bonavista Bay. The exposures at Stop 2-5 lie approximately 75 km south of Jenness' type section.

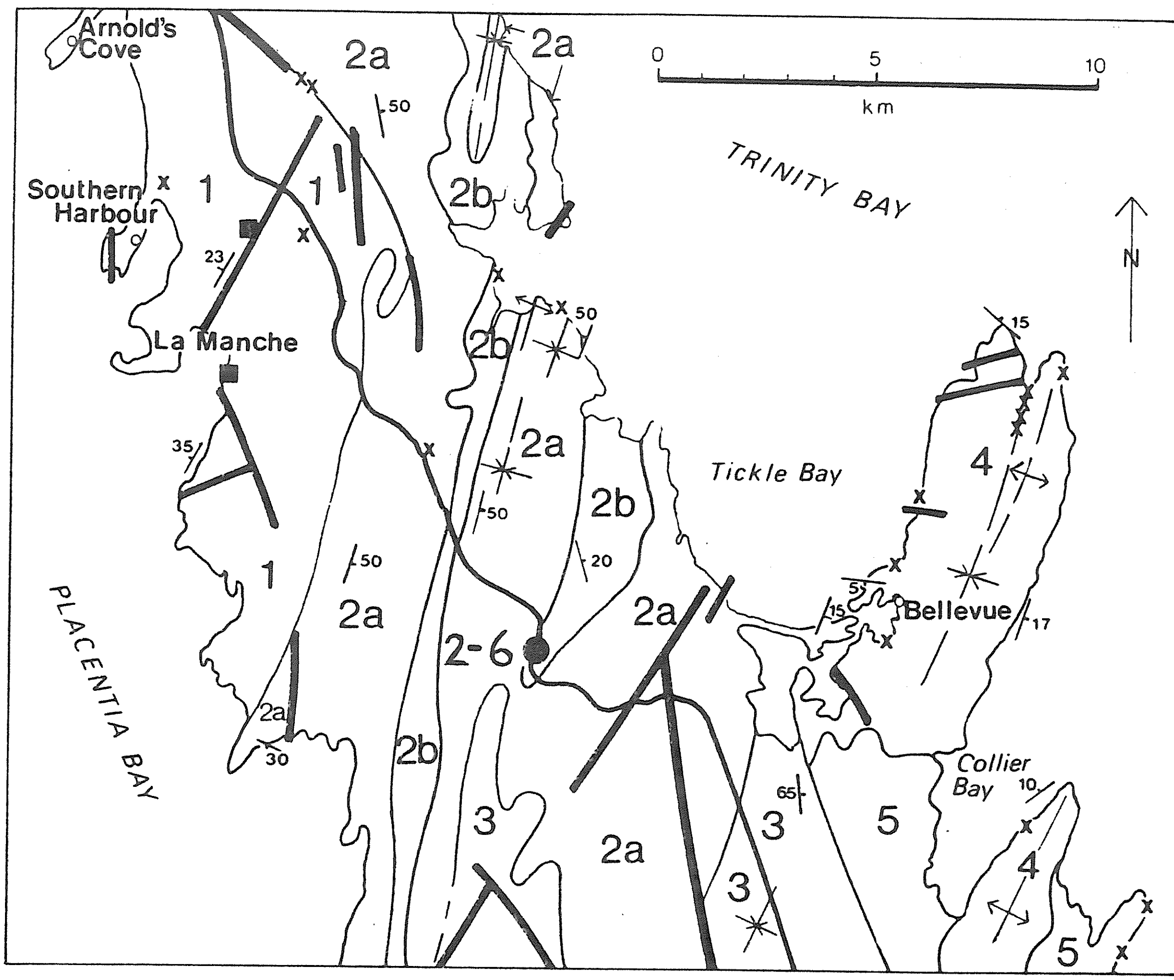
The Love Cove Group is now generally considered to be the oldest regional stratigraphic unit of the north western Avalon Zone (Hussey, 1979; O'Brien et al., 1984; O'Brien, 1987) and has been dated isotopically at 590 +/- 30 Ma and 607 +/- 25 Ma (Dallmeyer, 1980; Dallmeyer et al., 1981a). Although it has historically considered to be at least in part equivalent to the Harbour Main Group of the Avalon Peninsula, exact correlations are yet to be proven. In the southwest Bonavista Bay region, O'Brien (1987) and O'Brien and Knight (1988) have demonstrated that the group occupies a stratigraphic position conformably beneath the Connecting Point Group. The volcaniclastic sediments that occur at the top of the Love Cove Group and in the overlying Connecting Point Group possess the characteristics, composition and lithostratigraphic signature of volcanic arc basins (Knight and O'Brien, 1988). The volcanic rocks of the Love Cove Group in this area are mainly silicic pyroclastic and these were deposited together with tuffaceous clastic rocks in both subaqeous and subaerial environments. Compositions of volcanic rocks within the group range from basalt through andesite to rhyolite. In the region to the north of here, the group displays clear calc-alkaline chemical affinities; andesites are present but these are subordinate to all other rock types. Throughout much of its outcrop area, the group is intruded by calc-alkaline hornblende - biotite granitoids that are considered to be comagmatic with the volcanism (e.g. Dallmeyer et al., 1981a).

Typical exposures of the lower parts of the Love Cove Group will be examined at this stop. In a quarry on the west side of the highway, penetratively cleaved felsic lithic-crystal and lithic-crystal-vitric tuffs and fine grained tuff-breccias are exposed and can be examined on the weathered parts of the northeast quarry face. Pyroclastic fragments are flattened in the plane of a strong SL fabric. It is possible that at least some of this flattening may be a primary depositional feature. These pyroclastics are cut by chloritized metabasic dykes that are oriented subparallel to cleavage and altered to chlorite schists. The main deformation and metamorphism of the Love Cove Group in this area has been interpreted to be the results of Acadian orogenesis (Dallmeyer et al., 1983). Similar fabric intensities are seen in nearby Cambrian rocks. This regional fabric is truncated by the posttectonic (360 Ma) Ackley Granite. Recent work to the north, in Bonavista Bay (e.g. O'Brien, 1987; O'Brien and Knight, 1988) has documented existence of Precambrian fabrics in the Love Cove Group, and has rekindled the controversy surrounding the age of the first regional deformation of this group.

Stop 2-6: Volcanic rocks of the Musgravetown Group (Bull Arm Formation) (Figure 27)

This outcrop is located on the Trans-Canada Highway (Route 1) at the Doe Hills, Isthmus of Avalon. There are two possible localities to visit at this stop. The following discussion incorporates both. The isthmus is renowned for its foggy conditions and visibility is often near zero in this area. Exercise extreme caution in foggy weather: keep well away from the pavement!

Volcanic rocks of the Bull Arm Formation of the late Precambrian Musgravetown Group are exposed at both localities in this stop. The Musgravetown Group in this region is separated from the exposures on the western



LEGEND

LOWER CAMBRIAN

5 Sedimentary rocks

LATE PROTEROZOIC

4 Hodgewater Group sedimentary rocks
3 Musgravetown Group sedimentary rocks
2 Bull Arm Group: a, intermediate volcanic and associated sedimentary rocks; b, felsic volcanic and associated sedimentary rocks
1 Connecting Point Group sedimentary rocks

SYMBOLS

Fault
x Barite occurrences
■ Barite and sulphides
Trans Canada Highway
● Field-trip stop

Figure 27. General geology of the Isthmus of Avalon (Howse and Maloney, 1984).

Avalon by the Paradise Sound Fault (O'Brien and Taylor, 1983) and the intervening basin of late Precambrian Connecting Point Group sedimentary rocks. The Bull Arm Formation in this area consists of basalt, rhyolitic flows and ash flows and associated volcanogenic sediments that together have an aggregate thickness in the order of 2 to 2.5 km. The nature of the contact with the underlying Connecting Point Group in this area is somewhat enigmatic, but has been interpreted as conformable by McCartney (1967). In the region north of here, the contact is everywhere an angular unconformity (Hayes, 1948).

Red and buff rhyolite lava is exposed in the core of the Doe Hills anticline, a regional structure of apparent Acadian age (Hughes and Malpas, 1971). Flow banded and autobrecciated rhyolite is exposed in the older roadcuts on the s-shaped bend in the highway; in the newer, straight road cut, the same rhyolite is overlain by pyroclastic rocks and interbedded sediments on the east limb of the fold. Minor faults occur near the contact. In both localities the volcanics are cut by veins of white and pink tabular and bladed barite crystals. Sharp-bounded, barite veins are exposed in the older roadcut. Veins varying in thickness between 1 and 30 cm occur along joints that dip vertically and trend and trend from 010 and 030 degrees. Veinlets of barite and quartz, with or without specular hematite, with diffuse and, in places, sharp margins are exposed in the newer roadcuts; in both areas barite covers joint surfaces with attitude 130 to 160 degrees. The rhyolites in this area have suffered locally intense potash metasomatism, and contain up 10 percent K2O and 12 percent CaO; $K2^{O}$ content is higher near fractures (Hughes and Malpas, 1971). The feldspar in the rhyolites at this locality is a turbid untwinned orthoclase formed by the replacement of plagioclase. However, the unaltered lavas in the Bull Arm Formation in this region are characterized by relatively high $K2^{O}$ values, compared to other volcanic units.

The Bull Arm Formation represents an episode of volcanicity that is temporally and chemically distinct from the main pulse of Avalonian igneous activity reflected in the Harbour Main Group and Holyrood Granite of the type area on the Avalon Peninsula. Detailed studies and comparisons of the various belts of volcanic rocks throughout the central and western Avalon Zone are presently ongoing (O'Brien, 1987, O'Brien and Knight, 1988, Hayes, pers. comm., 1988) and the exact relationship amongst these various areas is presently uncertain. The stratigraphically equivalent volcanic succession in southwest Bonavista Bay and in the type area around Clode Sound is characterised by a bimodal succession of alkali basalt and rhyolites that display a continuum of compositions from alkali rhyolite to pantellerite and commendite (Hussey, 1979; O'Brien et al., 1986; O'Brien, unpublished data). These compositions appear to be a signature of the volcanic successions that postdate the earlier of two pulses of Avalonian orogeny documented in this region.

Hughes and Malpas (1971) suggested that the barite veins exposed at this stop were related to the metasomatism of the rocks, and resulted from exsolution of Ba from the feldspar. The alteration is apparently concentrated in the core of an Acadian fold, therefore they considered the metasomatism to be of similar age as the folding. Alternate hypotheses include: 1) the barite is the result of epithermal alteration associated with nearby Devonian intrusions that may even occur at depth at this locality, and 2) the metasomatism is a deuteric feature, the barite is a secondary feature.

END OF TRIP - RETURN TO ST. JOHN'S