GEOLOGICAL SURVEY OF CANADA

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Mineral deposits of New Brunswick and Nova Scotia (Field Trip 2)

edited by

D.R. Boyle

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MINERAL DEPOSITS OF NEW BRUNSWICK AND NOVA SCOTIA

[FIELD TRIP 2]

EDITED BY

D.R. BOYLE

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8TH IAGOD SYMPOSIUM

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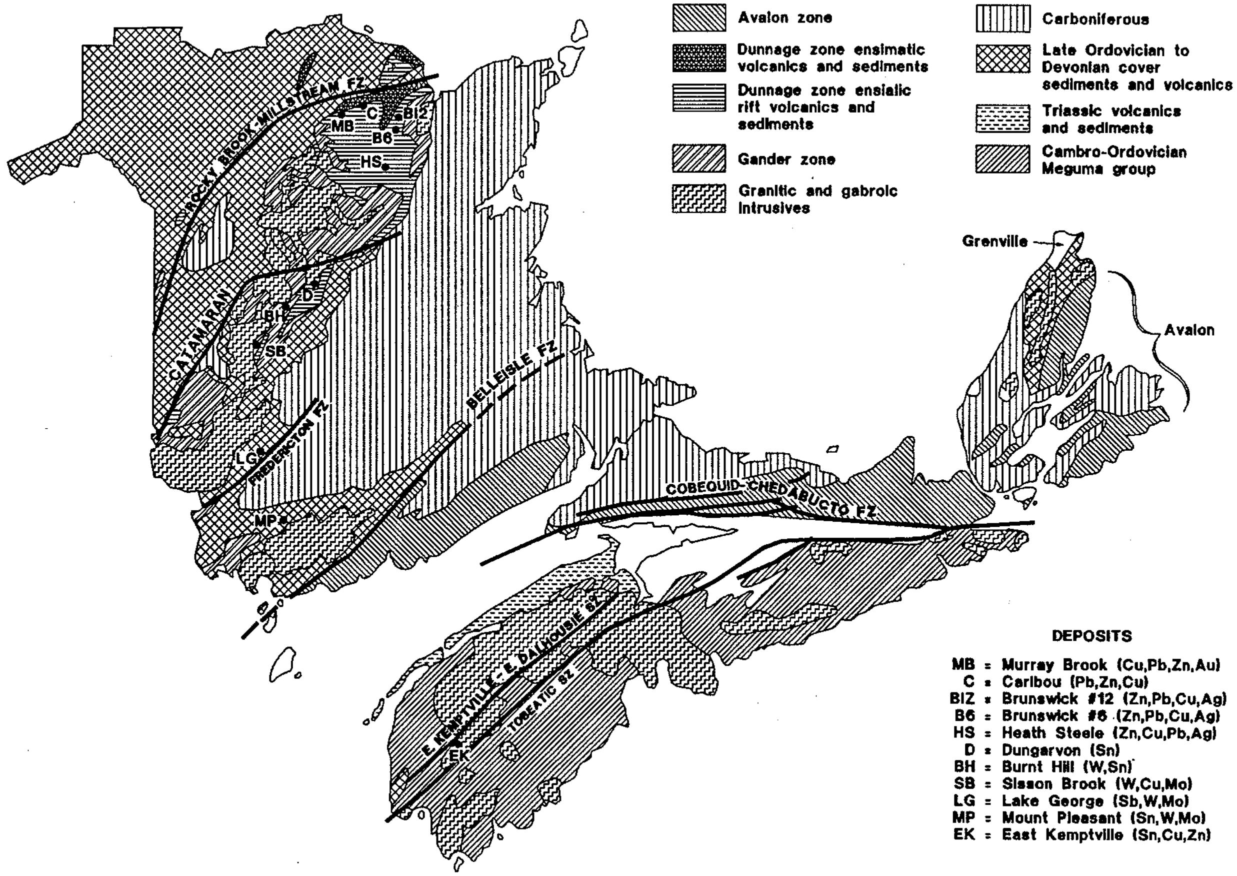

Figure 1. Generalized geology of New Brunswick and Nova Scotia showing main tectonic zones and locations of mineral deposits included in fieldtrip.

GEOLOGY OF ORDOVICIAN MASSIVE SULPHIDE DEPOSITS AND THEIR HOST ROCKS IN NORTHERN NEW BRUNSWICK

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GEOLOGICAL SETTING OF NORTHERN NEW BRUNSWICK

New Brunswick forms part of the northern Appalachians (Fig.1) and comprises three of the five tectono-stratigraphic zones recognised in this part of the orogen (Williams, 1979). These are from south to north respectively the Avalon, Gander and Dunnage zones (Fig. 1), which are mainly defined on their pre-Silurian geology. The Gander and Dunnage zones are commonly combined into the Central Mobile Belt (CMB) and are generally thought to represent respectively the vestiges of a Lower Paleozoic west facing passive margin and an oceanic (Iapetus) or back-arc basin (Iapetus 2) (Williams,1979; van Staal, 1987).

ORDOVICIAN TECTONOSTRATIGRAPHIC FRAMEWORK OF NORTHERN NEW BRUNSWICK

The Miramichi Highlands and the Elmtree-Belledune Inlier (Fig.2) are the principal areas where Cambro-Ordovician rocks of the Gander and Dunnage zone are exposed in northern New Brunswick.

The Cambro-Ordovician rocks in these areas have been separated into four groups: 1) the Miramichi Group, 2) the Tetagouche Group, 3) the Fournier Group and 4) the Balmoral Group.

The Balmoral Group, which contains Middle Ordovician andesitic and picritic volcanics and Caradocian black shale occurs in the Popelogan Inlier (Philpott, 1988) and is not treated here. The Miramichi Group (Fig. 2) comprises a monotonous sequence of quartz wacke and pelite of unknown thickness. Pelite becomes more abundant and graphitic towards the stratigraphic top, which is defined by the contact with the disconformably overlying volcanic and sedimentary rocks of the Tetagouche Group. The Miramichi Group is Arenigian and older (Fyffe et al., 1983) and defines the Gander Zone in northern New Brunswick.

The Tetagouche Group consists mainly of a voluminous suite of Middle Ordovician mafic and felsic volcanic rocks (Fig.3). Felsic volcanic rocks dominate and have compositions that range from dacite to rhyolite (Whitehead and Goodfellow, 1978; Winchester and van Staal, 1988).

The felsic volcanic rocks comprise a heterogeneous mixture of flows, shallow intrusions (e.g. porphyries), pyroclastic and proximal epiclastic deposits (van Staal, 1987). For mapping purposes, these rocks are generally divided into aphyric or feldspar-phyric rhyolite of the Flat Landing Brook Formation, and quartz and feldspar-phyric flow, pyroclastic and proximal epiclastic rocks of the Nepisiquit Falls Formation (cf. Skinner, 1974). Field, geochemical and petrographic studies have indicated that a large proportion of the felsic volcanics previously interpreted as ash flows represent rhyolite flows (van Staal, 1987; McCutcheon et al., 1989). The large areal extent of the rhyolite flows indicates that the felsic magma was relatively fluid, probably because it was dry and hot.

The felsic volcanic rocks are locally interbedded with thin, but generally laterally extensive bodies of iron formation, jasper, and a multicoloured (red, purple, green and black) Fe/Mn-rich phyllite. These metalliferous sediments are closely associated with most of the major base metal Zn-Pb-Cu-Ag massive sulphide deposits in northern New Brunswick. Locally the felsic volcanics are also interbedded with minor bodies of tholeiitic basalt. Enrichment in light

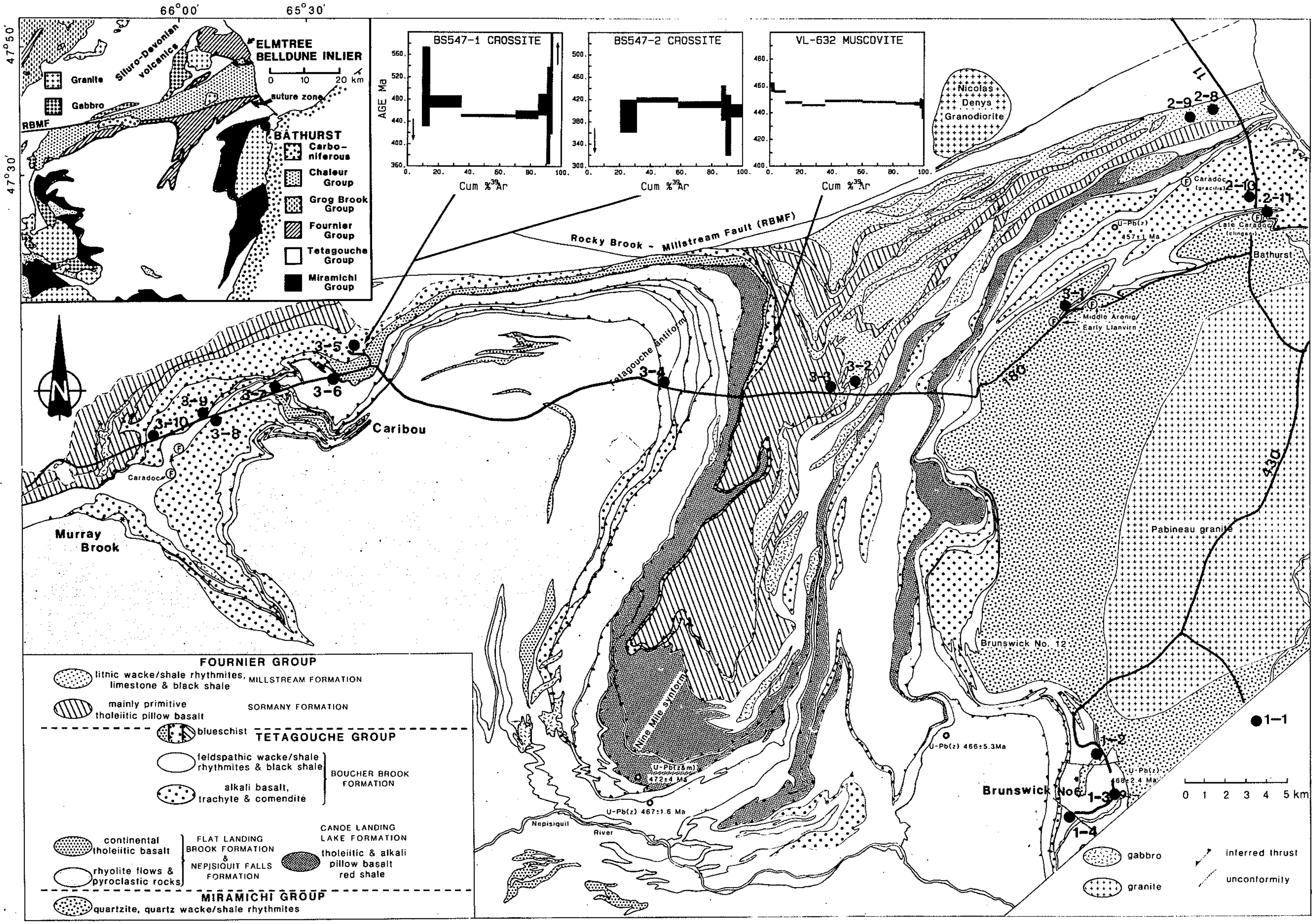

Figure 2. Simplified geology of the northern Miramichi Highlands with road tours.

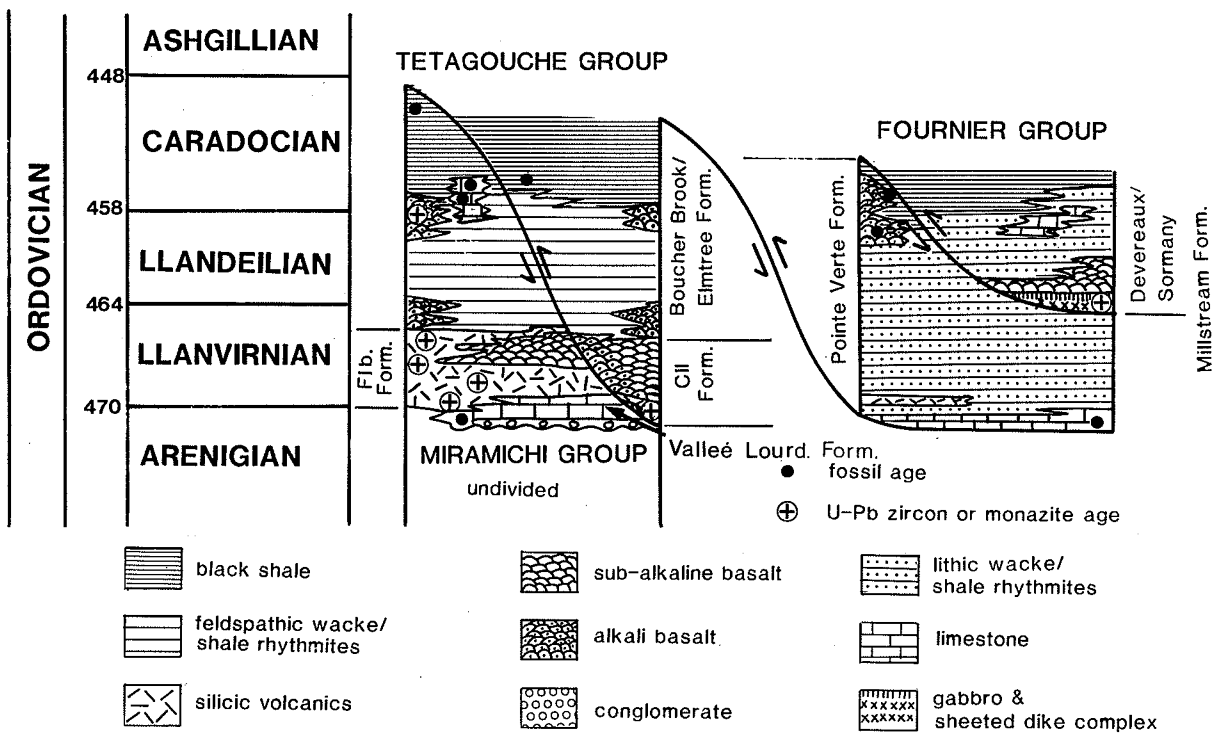

Figure 3. Generalized tectonostratigraphy of Ordovician rocks in the northern Miramichi Highlands and Elmtree-Belledune Inlier.

rare earth elements indicates a continental within plate setting. These basalt bodies are mainly massive flows, sills and pyroclasic tuffs, breccias and agglomerates. Pillows, common in all other basalt suites in the northern Miramichi Highlands, are rare or absent, suggesting a subaerial environment of deposition for at least part of these rocks. However, a shallow water environment of deposition is indicated by fossils and sedimentary structures for most of the associated iron formation and epiclastic, tuffaceous sediments. These include stromatolites and ooids in carbonate facies iron formation (McMillan, 1969; van Staal, 1987) and brachiopods and pelecypods in shaly sediments (Bolton, 1968; Fyffe, 1976; Gummer et al., 1978; Neuman, 1984). Part of the volcanic complex was probably emerged or experienced differential uplift during part of the volcanic history. Of interest are the rare pelecypods that have been found only in tuffaceous and/or epiclastic sediments that host the Devils Elbow (Bolton, 1968) and Taylor brook (Gummer et al., 1978) massive sulphide deposits. Analogous to the faunas found around hydrothermal vents on the seafloor, these pelecypods may represent fossilized Ordovician vent faunas.

U-Pb zircon ages of the felsic volcanics range from 472 to 466 Ma. (Sullivan and van Staal, 1989), and cluster around 468 Ma.. In the northeastern part of the northern Miramichi Highlands the felsic volcanic rocks are conformably underlain by a thin unit of shallow water arenaceous or rudaceous limestone and/or a calcareous phyllite. This rock unit has been named the Vallee Lourdes Formation (Van Staal et al., 1988b). Brachiopods and conodonts indicate a middle Arenigian to early Llanvirnian age for this formation. These ages suggest that the formations containing the felsic volcanic and minor interbedded sedimentary rocks are mainly Llanvirnian in age. The Vallee Lourdes Formation defines the base of the Tetagouche Group (Fig. 3) and can be seen to lie disconformably on top of the Miramichi Group during low water levels in the Tetagouche river, east of Tetagouche Falls, 12 km. west of the city of Bathurst. The disconformity is marked by a thin bed of conglomerate, which contains quartzite and shale pebbles of the underlying Miramichi Group.

This conglomerate is interpreted to mark a bulging disconformity that formed as a result of back-arc rifting during the middle to late Arenigian.

The Flat Landing brook and Nepisiquit Falls formations are conformably overlain by the Boucher Brook Formation (Fig. 2 and 3), which contains thin bedded feldspathic wacke/shale rhythmites, black shale and a chemically distinct (low-Cr) alkali basalt with minor trachyandesite, trachyte and comendite. The Boucher Brook Formation ranges in age from the Llandeilian to latest Caradocian on basis of several fossil localities and U-Pb zircon ages of the volcanics (Nowlan, 1981; Riva and Malo, 1988; van Staal et al. 1988b; Sullivan and van Staal, 1989). The Elmtree Formation in the Elmtree-Belledune Inlier contains lithologically and chemically similar rocks as the Boucher Brook Formation (Fig. 1, 3 and 4).

The Flat Landing Brook/Boucher Brook package is structurally overlain by another volcanic unit/Boucher Brook package (Fig. 2 and 3). The volcanic unit consists mainly of tholeiitic- and minor alkali pillow basalts of the Canoe Landing Lake Formation, which are also Llanvirnian in age (van Staal and Sullivan, unpubl. res.). The presence of interbedded red Fe/Mn-rich phyllites supports the age dating. Since old overlies young the contact between these two packages is interpreted as a major thrust, which is marked by the presence of a narrow zone of phyllonite (van Staal, 1986). Each of these two tectonostratigraphic packages is internally imbricated (Fig. 2) and the thrust zones are marked by cut-offs and, where exposed, by zones of phyllonite or mylonite.

The Tetagouche Group is structurally overlain by the Fournier Group, which consists of the Sormany- and Millstream formations in the northern Miramichi Highlands (Fig. 3). The Sormany formation consist of pillow basalts and minor gabbro. The basalts are mainly primitive tholeiites with MORB-like compositions but also show compositions intermediate between MORB and IAT. These basalts are chemically and lithologically equivalent to part of the ophiolitic Devereaux Formation (Pajari et al., 1977) in the Elmtree-Belledune Inlier. This part of the Sormany Formation is therefore also interpreted as a fragment of back-arc oceanic crust. A U-Pb zircon age of 463.9 +/- 1 Ma. for a pegmatitic gabbro pod in the gabbroic part of the Devereaux Formation indicates that formation of oceanic crust was slightly later than eruption of the majority of the rift volcanics and supports the back-arc setting proposed by van Staal (1987). The Millstream Formation consists of lithic wacke/shale rhythmites, minor conglomerate, arkose or feldspathic wacke, limestone, (high-Cr) alkali basalt and black shale. Lithologically equivalent rocks are present in the Pointe Verte Formation of the Elmtree-Belledune Inlier, which ranges in age from middle/late Arenigian to early Caradocian (Nowlan, 1983, 1988a; Riva in Fyffe, 1986). The Devereaux- and Pointe Verte Formations define the Fournier Group in the Elmtree-Belledune Inlier (Fig. 3 and 4). The contact between these two rock formations is a thrust (Fig.4) since the Devereaux Formation is older than the underlying, early Caradocian part of the Pointe Verte Formation (Fig. 3). A narrow zone of amphibole-chlorite phyllonite marks the thrust in the field. The contact between the Pointe Verte Formation and the structurally underlying Elmtree Formation of the Tetagouche Group in the Elmtree-Belledune Inlier is also interpreted as a thrust. The contact between these two units is marked by a narrow zone of phyllonite and melange exposed in the the Elmtree river where it separates late Caradocian black shales of the Elmtree Formation (Dean, 1975) from structurally overlying middle to late Arenigian limestone (Nowlan, 1988a) of the Pointe Verte Formation.

The contact between the Fournier and Tetagouche Groups is a major thrust zone along its entire length (Fig. 2, 3 and 4) and can be traced from the Miramichi Highlands into the Elmtree-Belledune Inlier with an apparent dextral offset along the RBMF. An extensive belt of sodic amphibole bearing blueschists, unparalleled in the Appalachian/Caledonian orogen, defines this contact for at least 70 km. in the northern Miramichi Highlands, suggesting that this contact marks a suture.

STRUCTURE AND TECTONICS

The Ordovician rocks in northern New Brunswick have undergone complex polyphase folding and faulting (van Staal and Williams, 1984; Van Staal, 1987). At least 5 generations of folds have been demonstrated the on basis of overprinting relationships. The earliest structures comprise a strong layering parallel foliation (S1), asymmetrical intrafolial folds (F1) and a stretching lineation (L1). The D1 structures are typically concentrated in narrow zones of high strain, which commonly coincide with repetitions in stratigraphy. The D1 structures are interpreted as a result of a progressive deformation associated with thrusting. The D1 deformation is markedly heterogeneous on all scales and consequently the amount of strain recorded in the rocks as well as the spectrum of structures present varies from place to place. The narrow zones of rocks affected by D1 are strongly altered and transformed into phyllonites or mylonites. This reaction and fabric softening of the rocks localizes the subsequent increments of the polyphase deformation and enhances the heterogeneous distribution of the strain.

The D1 structures are refolded by F2 into tight to isoclinal folds (van Staal and Williams, 1984) that define flat and steep belts (van Staal, 1987). The coherence of the D1 tectonostratigraphy was retained after F2 folding in the northernmost part of the Miramichi Highlands (Fig. 2) suggesting that the F2 enveloping surface makes a small angle with the orientation of S1 in this part of the area. This feature and several other criteria suggest that this part of the Miramichi Highlands lies on the northern limb of a regional scale F2 antiform cored by the felsic volcanics. The F2 structures are refolded by open recumbent F3 folds, which tend to be symmetrical and restricted to outcrop scale (de Roo et al., 1990) indicating a small degree of vertical shortening. The D1, F2 and F3 structures are overprinted and refolded by F4 and F5 folds and kinks which range in size from mm.- to km. scale. The Pabineau structures (van Staal and Williams, 1984) are F4 folds whereas the Nine Mile synform and Tetagouche antiform (Fig. 2) are F5 structures (van Staal, 1986, 1987). The steep northerly plunge (60-70°) of these structures created a large amount of structural relief in the core of the Nine Mile synform. Based on a lack of overprinting relationships, F3 was previously thought to represent F5 (van Staal and Williams, 1984). As a consequence of recent work (de Roo et al., 1990) the nomenclature has changed and F3 and F4 have become F4 and F5 respectively. F4 and F5 are interpreted as a result of dextral transpression that culminated in a

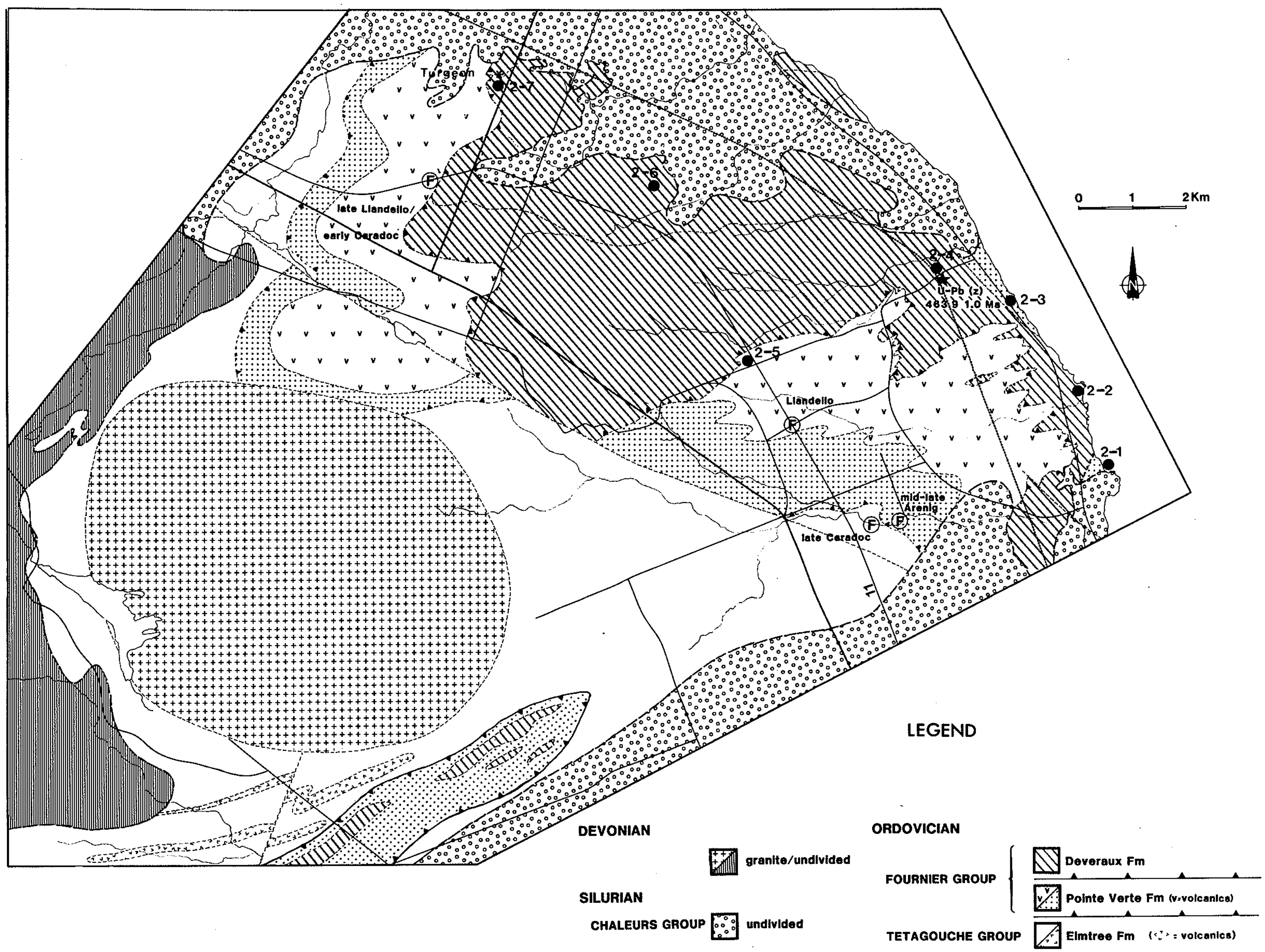

Figure 4. Geology of the Elmtree-Belledune Inlier.

large dextral offset along the Rocky Brook Millstream fault (RBMF) (van Staal and Langton, 1988). A minimum displacement of 20 km. is indicated by the offset of Siluro-Devonian gabbros and a displacement of ca. 50 km. is suggested by the offset of the Fournier/Tetagouche Group suture.

Early plate tectonic models suggested that the Tetagouche rocks represented the remnants of the Taconic arc formed above an eastward dipping subduction zone (e.g. Pajari et al., 1977). Chemical compositions of volcanic rocks of the Tetagouche Group (see above) do not correspond with a magmatic arc setting but instead closely resemble volcanics found in an ensialic rifting environment (van Staal, 1987; Winchester and van Staal, 1988). Van Staal (1987) therefore proposed that the bulk of the volcanic rocks of the Tetagouche Group formed in a Taconic back-arc basin that started to open in middle to late Arenigian times. The Devereaux and Sormany formations of the Fournier Group represent back-arc basin oceanic crust. This hypotheses was tested by age dating which showed that the oceanic crust is slightly younger than the rifting suite of the Tetagouche Group (see above) and thus consistent with the back-arc basin model. Restoration of seismically defined lower crustal blocks to a precollisional configuration also supports the presence of a wide back-arc basin (Stockmal et al., 1990).

The back-arc basin (Iapetus 2, van der Pluijm and van Staal, 1988) started to close in Late Ordovician times by northwards directed subduction (van Staal, 1987), which lasted at least until Late Silurian times. This time period is constrained by 1) Ar^{39}/Ar^{40} age dating of crossite and phengite from the blueschist belt, which yielded ages ranging from 450 to 410 Ma. (Ravenhurst et al., 1990); 2) the youngest rocks of the Tetagouche Group involved in the D1 thrusting are late Caradocian in age (Riva and Malo, 1988); and 3) the blueschist belt is unconformably overlain by Late Silurian (Ludlovian) conglomerates of the Chaleur Group (Helmstaedt, 1971). Within this tectonic scenario, D1 and M1 are seen as the products of the subduction-related deformation and metamorphism. Post-D1 ductile deformation resulted mainly from the oblique collision between North America (with the accreted Taconic arc) and Avalonia.

MASSIVE SULPHIDE DEPOSITS

The Tetagouche Group in the northern Miramichi Highlands is characterised by an anomalous abundance of base metal-rich massive sulphide deposits. Over 30 mineral deposits are known (Davies, 1979). At present ore is produced from the Brunswick No. 12 and Heath Steele mines. The Brunswick No. 6, Wedge and Caribou deposits have been mined in the past for base metals. Economic concentrations of gold and silver in the gossans overlying the massive sulphides have been mined in the Caribou and Heath Steele mines and are at present extracted from the gossan of the Murray Brook deposit (Fig. 2). The large iron formation in the hanging wall of the Austin Brook deposit (Fig. 5) was mined for iron in the beginning of this century (Boyle and Davies, 1964).

The massive sulphide deposits of the Tetagouche Group are invariably closely associated with the felsic volcanic and epiclastic rocks of the Nepisiquit Falls and Flat Landing Brook formations. At least two, and possibly three sets of deposits can be recognised (van Staal and Williams, 1984; van Staal, 1986). The first set of deposits, what is referred to as the Brunswick-type, forms an integral part of a laterally extensive Algoma-type iron formation (McAllister, 1960; Davies, 1972). Included in this set are the Brunswick No.12 (Luff, 1975), No.6 (Boyle and Davies, 1964), Austin Brook (Boyle and Davies, 1964; Davies, 1972), Flat Landing Brook (Troop, 1984), Key Anacon (Saif et al., 1978) and Heath Steele (Whitehead, 1973) deposits. The Brunswick-type occurs generally along or close to the contact between the Gordon Meadow Brook Formation of the Miramichi Group and the Nepisiquit Falls Formation of the Tetagouche Group. Since rocks of the latter formation have a U-Pb age of 468 Ma these deposits formed in Llanvirnian times, during the earliest stages of felsic volcanism. The second set of massive sulphide deposits, the Caribou-type, is generally hosted by feldspathic wackes and phyllites of the Boucher Brook Formation or occurs at the contact with the underlying felsic volcanics of the Flat Landing Brook or Nepisiquit Falls Formation (McAllister, 1960; Helmstaedt, 1973; van Staal, 1986). The Caribou-type thus seems to occur at a higher stratigraphic level than the Brunswick-type near the endstages of felsic volcanism. This set includes the Nepisiquit A,B,C, Nine Mile Brook, Canoe

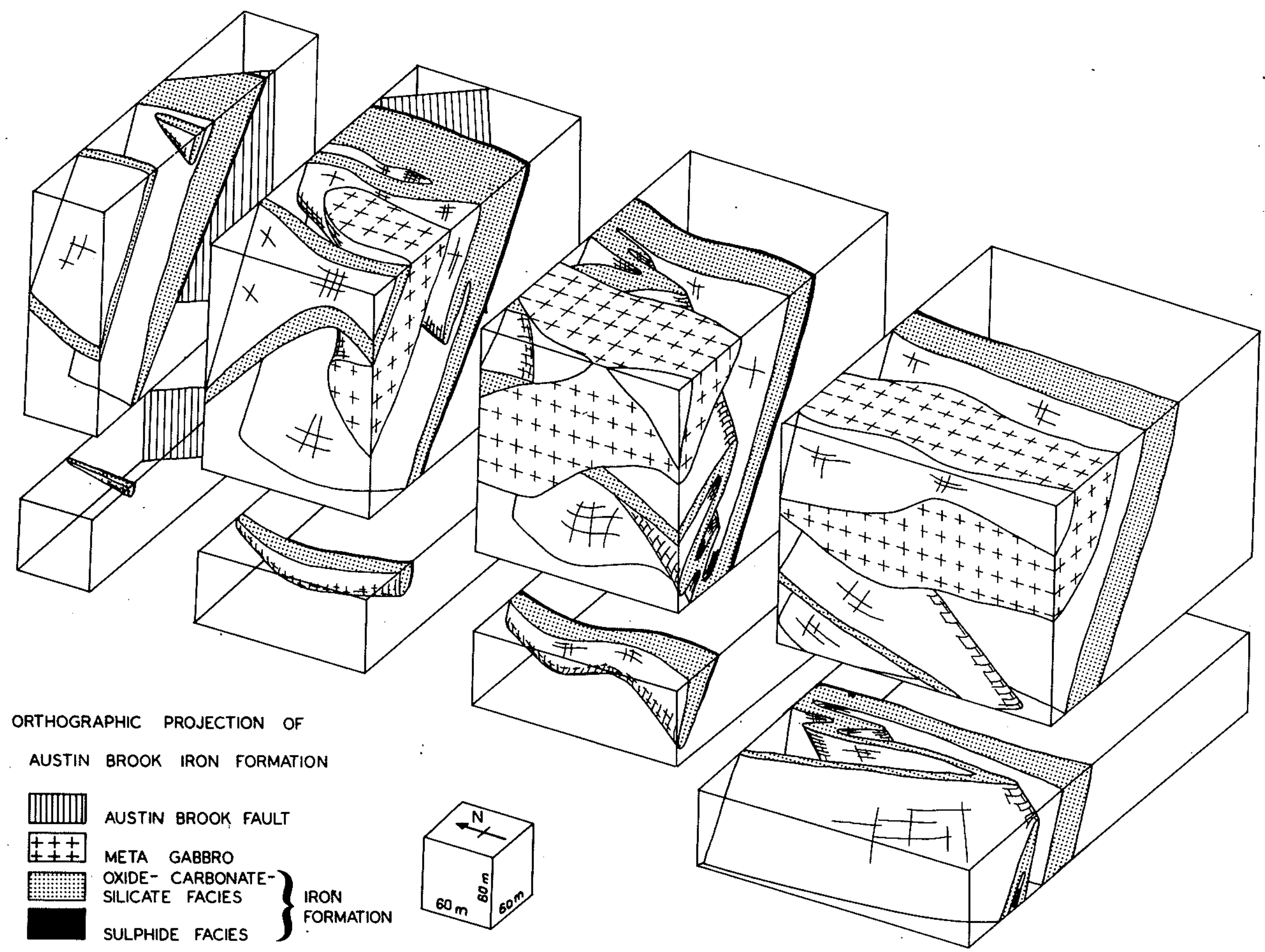

Figure 5. Block diagram of the Austin Brook iron formation body (after Van Staal, 1987).

Landing Lake, Orvan Brook, Murray Brook and the Caribou deposits (Fig. 2). The Caribou-type does not contain a laterally extensive Algoma- type iron formation, although it is generally associated with a red, relatively Fe/Mn-rich shaly and cherty phyllite that approximately occupies the same stratigraphic position as the sulphides distal from the massive sulphide deposits (van Staal, 1986). However, the sulphides and the red phyllite have not been found in contact or in close proximity with one another.

This twofold division is supported by as yet unpublished (but referred to in Swinden and Thorpe, 1984) lead isotope data, which show that the Brunswick type galenas have slightly lower 206Pb/204Pb and 208Pb/204Pb ratios than the galenas of the Caribou type. This division is also consistent with sulfur isotope patterns as compiled by Franklin et al. (1981).

A tentative third set of massive sulphide deposits, the Halfmile Lake-type, is hosted by silty and shaly phyllites of the Gordon Meadow Brook Formation close to or at the contact with the overlying rocks of the Nepisiquit Falls Formation. This set includes small deposits such as the Halfmile Lake, Chester (Harley, 1979; Jambor,1979) and the FAB (van Staal and Williams, 1984). Their overall stratigraphic position thus appears to be slightly lower than the Brunswick-type, although locally the latter is also in contact with the Gordon Meadow Brook Formation. However, the Halfmile Lake-type deposits have no associated iron formation-like metalliferous sediments and we therefore tentatively treat them as a separate type.

The three types of deposits range in size from small showings to supergiants such as the Brunswick No. 12 deposit and consist of concordant massive and disseminated bodies of pyrite, sphalerite, galena, chalcopyrite, magnetite and in places pyrrhotite. Several other sulphides, particularly arsenopyrite, sulphosalts, and oxides occur in minor amounts.

The large deposits generally display a large scale mineralogical and chemical zonation as well a small scale compositional layering or banding (Boyle and Davies, 1964; Rutledge, 1972; Luff, 1977; Jambor, 1979). The zonation is best developed in the Brunswick-type, which exhibits variation perpendicular and parallel to strike. The lateral zonation in the Brunswick-type is defined by a gradual change from massive sulphide into iron formation (Boyle and Davies, 1964; Luff, 1975; Troop, 1984; van Staal, 1985), such that there is locally a mixed sulphide-iron formation body. In the Caribou deposit lateral zoning is illustrated by a decrease in magnetite, chalcopyrite and Zn,Pb,Ag from the west limb of the Caribou fold, around the nose to the east limb (Jambor, 1979). The vertical zonation in the Brunswick-type sulphide deposits is ideally made up of four zones (Rutledge, 1972; Luff, 1977; van Staal and Williams, 1984). These are: 1) a massive or crudely layered pyrite body with variable amounts of pyrrhotite, magnetite and chalcopyrite at the footwall; 2) a zone of well layered (cm.-mm. scale) pyrite, sphalerite and galena with minor chalcopyrite and pyrrhotite; 3) a massive pyrite body with thin discontinuous layers or lenses of sphalerite and galena and 4) iron formation.

There is thus a decrease in the Cu-content between zone 1 and 3 and an apparent enrichment of Zn and Pb in the zones 2 and 3. This zonation is best developed in the Brunswick No.6 and No. 12 and parts of the Heath Steele orebodies. The principles of this zonation are also present in the Austin Brook deposit (Fig.5),although the sulphide zonation is condensed here (van Staal, 1985). A decrease in Cu/Pb+Zn from stratigraphic footwall to hangingwall has also been observed in the Caribou and Half Mile Lake deposits (Jambor, 1979).

Although the sulphide zonation is not always continuous due to primary lateral impersistence of the zones and complications induced during deformation, it is interpreted as a pre-deformational feature (van Staal and Williams, 1984) that can be used as a younging indicator (c.f. Large, 1977; Stanton, 1979). For example the metal zonation in the Halfmile Lake deposit indicates that the orebody is tectonically inverted. Another type of zonation is displayed by the Brunswick-type iron formation on a regional scale. For instance that part of the iron formation that is continuous between the Austin Brook deposit and the Brunswick No. 12 mine changes respectively from an oxide iron formation consisting dominantly of banded hematite, magnetite and jasper to a carbonate and/or silicate iron formation.

A chlorite phyllite with variable amounts of disseminated pyrite-pyrrhotite-chalcopyrite (>0.5 % Cu) typifies the footwall of the Half Mile Lake type while the footwall of the Brunswick-type deposits generally comprise chlorite and/or sericite-chlorite schists, which locally contain relict quartz phenocrysts. The chlorite in the footwall of the Brunswick-type is typically iron-rich (Davies, 1972; Juras, 1981; van Staal, 1985) while pyrite, pyrrhotite, magnetite and apatite are locally important accessory minerals. These rocks represent, at least in part, altered and metamorphosed tuffites, but probably also contain a chemical component, sometimes referred to as the footwall iron formation (Jambor, 1979; van Staal and Williams, 1984). However, both the Brunswick No.6 and No.12 as well as the Heath Steele deposits are locally in direct contact with phyllites of the Miramichi Group (Luff, 1977; van Staal, 1985; Moreton and Williams, 1986), suggesting that these sulphide bodies form an integral part of the Nepisiguit Falls Formation. They appear to be conformably overlain by aphyric or feldspar phyric rhyolite and rhyolite lapilli or ash tuffs of the Flat Landing Brook Formation, although a very Mg-rich chlorite phyllite layer overlies the iron formation in the hangingwall of the Brunswick No.12 deposit (van Staal, 1985). Preserved fine scale bedding or laminations, but no other sedimentary structures, in low strain zones and some quartz phenoclasts suggest that the protolith of this phyllite was deposited under relatively quiescent conditions, probably having a tuffaceous or epiclastic component. Mg-rich chlorites were also observed in the hangingwall rather than the footwall of the Caribou deposit by Jambor (1979).

The Cu-pyrite Turgeon deposit is the only known sulphide body of significance in the Fournier Group. A small Cu-pyrite occurrence (Middle River) in the northern Miramichi Highlands also occurs in the Fournier Group. The Turgeon deposit, located near Belledune approximately 30 km northwest of bathurst (Fig. 4), is hosted by massive to pillowed tholeiitic basalt flows that are interbedded with jasper and wacke/shale rhythmites. Close to the sulphide mineralization the mafic volcanic rocks are silicified, chloritizised and locally brecciated (Kettles, 1987).

Most of the sulphide mineralization in the Turgeon deposit (pyrite, pyrrhotite, chalcopyrite and sphalerite) occurs in stringer-type or disseminated bodies (ca. 1.1 million tonnes grading 1.83% Cu and 1.26% Zn). Massive Zn-Pb-Cu pyrite lenses or pods (64 thousand tonnes grading 5.28%Zn, 1.15% Pb and 0.53% Cu) comprise a smaller fraction of the sulphide mineralization (Fyffe et al., in press). The sulphide zones plunge steeply to the southeast and crosscut the relatively shallow, east-dipping volcanic sequence that hosts the sulphide mineralization. These crosscutting zones possibly represent old fractures that channeled metal-bearing fluids during ocean floor related hydrothermal activity.

ROAD LOG FOR EXCURSION TO MIRAMICHI AND TETAGOUCHE GROUPS IN THE KEY ANACON AND BRUNSWICK MINES AREA. ROAD TOUR NO. 1

0 km Junction of Vanier Boulevard with Bathurst by pass (Highway No. 11, exit 310). Proceed south on Route No. 11.

6.2 km Take exit 304 and proceed southwards on Route 430.

22.6 km Turn east on Route 360 towards Allardville.

27.7 km
Stop 1 - 1

Nepisiguit River Bridge at the Key Anacon deposit Quartz wacke (locally feldspathic)/shale rhythmites of the Miramichi Group are in contact with rusty sericitic phyllites of the Nepisiguit Falls Formation of the Tetagouche Group (Fig. 6). These phyllites represent either highly strained and altered felsic tuffs or epiclastic rocks derived from a felsic volcanic protolith.

Cross-bedding and grading in the wacke beds are best preserved on the eastern side of the river and indicate that the Miramichi Group is older than the Tetagouche Group.

Strain is mainly concentrated in the shale beds with the competent wacke beds behaving as relatively rigid bodies that were in part boudinaged. This partitioning of the deformation is responsible for the preservation of the sedimentary structures in the thick, competent wacke beds. Close to the contact with the sericitic phyllites, the wacke/shale rhythmites grade into a thin layer of black, graphitic shale. Black shale is commonly found near the top of the Miramichi Group.

The vergence of the F2 folds changes across the outcrop and suggests the presence of a large , steeply plunging F2 fold, the hinge of which is obscured by faulting (Fig. 6). This structure was outlined by Saif et al. (1978) from mapping and drill hole interpretation.

The F2 structures fold a well developed differentiated layering (S1) but are themselves overprinted by NE-trending F4 and NNW to W-trending F5 folds and kinks. Recumbent kinks or open folds may represent F3 structures, although overprinting relationships with F4 and F5 to prove this tentative grouping on basis of style and orientation have not been observed.

An important massive sulphide deposit, the Key Anacon deposit (Saif et al., 1978) occurs on the eastern side of the river and is stratigraphically underlain by the sericitic phyllites and overlain by alkali basalts and sediments of the Boucher Brook Formation. These alkali basalts are chemically similar to those overlying the Brunswick No. 12 deposit and contain appreciable amounts of magnetite, which makes them a good magnetic marker. Where these basalts are strongly deformed into phyllonites, they can be mistaken for silicate iron formation (Saif, 1980).

0 km Return westward to road 430 and turn left.

6.7 km Turn left off the paved highway onto gravel road and proceed southwards to intersection with the road to Nepisiguit Falls on the left.

8.5 km Turn left on Nepisiguit Falls road.

10.5km
Stop 1 - 2 Knights Brook

Quartz wacke/shale rhythmites of the Miramichi Group are exposed to the right on Knight Brook. The rocks are strongly folded by F2, which are accompanied locally by slides parallel to bedding (Fig. 7). The F2 structures are overprinted by the S4 (trending N 50°E) and S5 (trending N 120°E) cleavages.

12.1 km
Stop 1 - 3 Nepisiguit Power dam

Cross the dam and descend to the rocks exposed on the eastern side of the dam.

Exposures show very clean and polished rocks of the Nepisiguit Falls Formation that forms the footwall to the Brunswick No.12 and No. 6 deposits. These rocks comprise the QAS (quartz augen schist), QFAS (quartz feldspar augen schist) and sericitic phyllites, previously interpreted as felsic pyroclastic rocks. The QFAS form massive, homogeneous bodies of quartz and feldspar phyric rhyolite, locally containing lithic clasts.

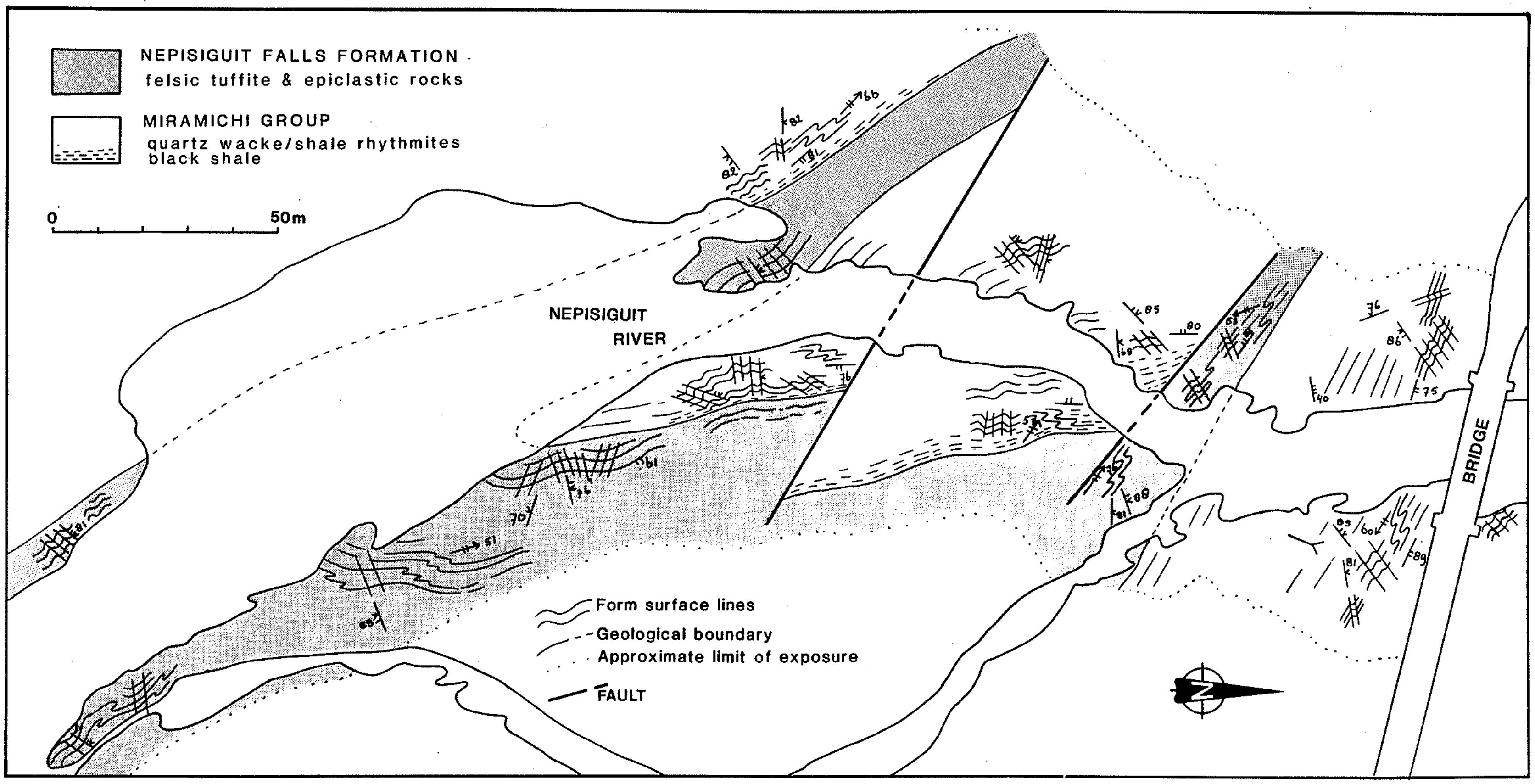

Figure 6. Geology and form surface map of the Key Anacon mine area.

The feldspar phenocrysts are very large (up to 1-2 cm) but not broken suggesting a lava flow or sill rather than a pyroclastic flow origin. The QFAS are surrounded by well layered and graded QAS which do not contain feldspar phenocrysts. The quartz augens are subrounded and do not contain embayments, which together with the sedimentary structures clearly indicates that they are sedimentary rocks (McCutcheon et al., 1989); either proximal epiclastic deposits or a mixture of pyroclastic and epiclastic material. Igneous zircons in the QAS yield similar ages to the zircons in the volcanics with which they are interbedded. This favours an origin as a mixture of pyroclastic and epiclastic material.

14.6 km Proceed westwards from dam for approximately 2.5 km to Austin Brook Quarry on the left side of the road.

Stop 1 - 4

The quarry contains a thick body of oxide iron formation that is folded into an isoclinal, moderately to shallowly south plunging S-shaped F2 fold (Fig.5). The iron formation is stratigraphically underlain by a lenticular massive sulphide body up to several meters thick. The sulphide body has a massive pyrite base followed in turn by a Zn/Pb rich (up to 17% combined) zone and a cherty pyrite layer respectively. The sulphides are in turn underlain by a thin layer (0-2m) of chloritic phyllites containing anomalous amounts of apatite. The chlorite is characteristically iron rich. The footwall of the metalliferous horizon comprises altered and highly strained pyrite bearing sericitic phyllites of the Nepisiguit Falls Formation. Except for the anomalously large oxide iron formation, the Austin Brook deposit is identical in make up to the Brunswick type deposits with which it is connected by a layer of metalliferous sediment (iron formation).

The iron formation shows abundant minor folding, comprising both F1 and F2 folds. These folds are interpreted to be post-lithification structures based on the following arguments: 1) the folds are coplanar to F1 and F2 folds developed in the surrounding volcanic rocks and also have the same style and plunge directions; 2) Quartz in jasper layers and intrafolial folded quartz veins shows evidence of intracrystalline deformation and grain boundary adjustment and has a c-axis fabric related to the folding; 3) Hematite is strongly foliated, kinked or bent in the hinges of the F1 and F2 folds, indicating intracrystalline deformation.

The great abundance of small scale folds in the iron formation with respect to the surrounding volcanic rocks is probably caused by the presence of a well developed compositional layering defined by alternating competent (jasper and magnetite) and incompetent (hematite) beds.

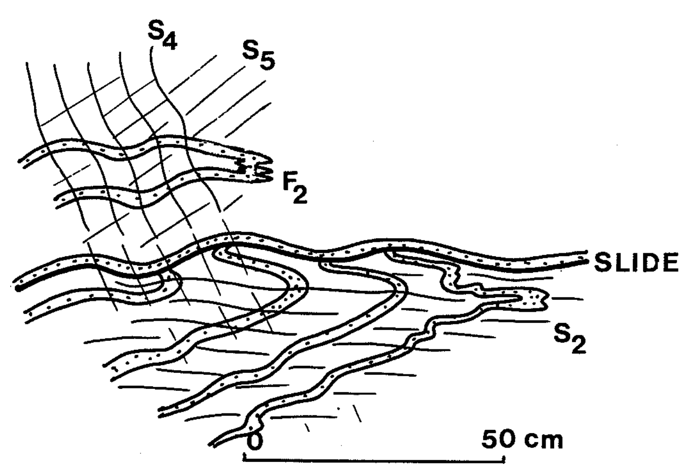

Figure 7. Overprinting relationships between S2, S4 and S5.

ROAD LOG FOR EXCURSION TO FOURNIER GROUP AND PARTS OF THE TETAGOUCHE GROUP IN THE BELLEDUNE-ELMTREE INLIER AND NORTHERNMOST PART OF MIRAMICHI HIGHLANDS. ROAD TOUR NO. 2.

0 km Exit 310 on Bathurst bypass (Rte. 11). Turn north onto bypass

16.3 km Turn right at Exit 326 onto Rue Laplante to Petit Rocher.

0 km Turn left at intersection with Rue Principale (Rte. 134). Restart road log.

4 km Turn right onto Doucet Road to the coast.

Stop 2 - 1 Limestone Point (Fig.8).

Middle Ordovician feldspathic (pinkish albitised plagioclase fragments) to lithic wackes are rhythmically interbedded with shale. The wacke beds are relatively thick and commonly graded. Note interbedded pinkish coloured, aphyric felsite beds (trachytic?) with micro laminations, which are strongly folded. Going southwards the wackes are gradually replaced by reddish coloured siltstones, which are tightly folded by shallowly to steeply eastwards plunging folds. The reddish siltstones are bounded to the south with a sharp nearly vertical contact by late Llandoverian limestone of the Armstrong Brook Formation (Nowlan, 1988).

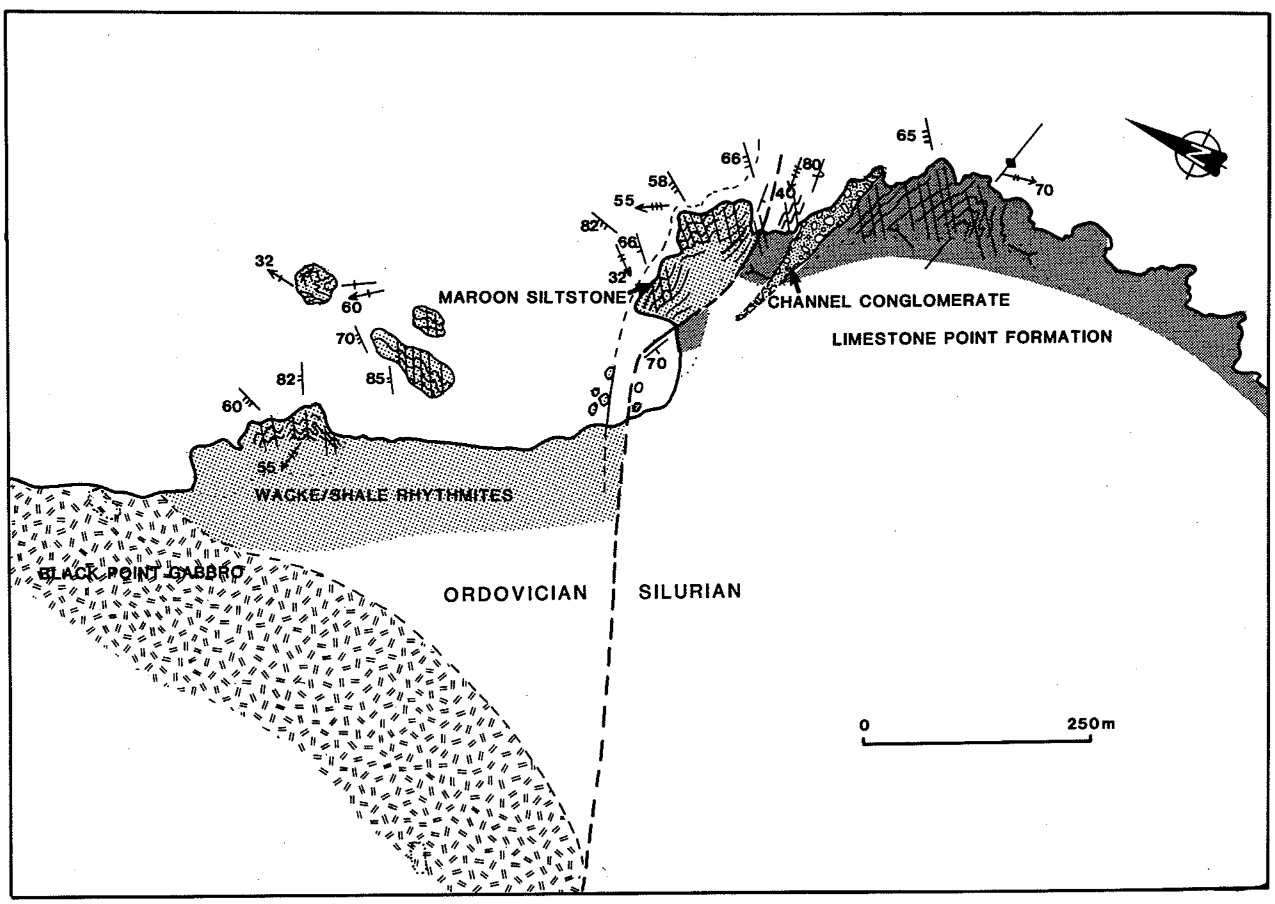

Figure 8. Geology and form surface map of the Limestone Point area.

Five meters south of this contact, a small channel of reddish conglomerate is interleaved with the limestone (Fig. 8). The erosional base of the channel appears to have acted as an inhomogeneity during deformation and localized the formation of asymmetrical folds. Basalt and gabbro pebbles, presumably derived from the ophiolitic Devereaux Formation indicate that the accretionary complex in which the ophiolitic fragment was incorporated was locally exposed and subject to erosion in Early Silurian times. The regional cleavage in the Silurian rocks transects the upright folds in an anticlockwise fashion (Stringer, 1975) indicating dextral transpression. This Silurian cleavage corresponds with the axial plane cleavage of asymmetrical, Z-shaped F3 folds in the Ordovivician rocks. The presence of polyphase folds and cleavages locally in the Silurian rocks indicates that the Silurian regional cleavage is also a third generation structure. The apparent hiatus between the Silurian and Ordovician rocks, combined with the sharp, conformable contact suggest the presence of a fault or a major disconformity.

0 km Return to Route 134 and turn north. Restart road log

1.6 km Turn right, park car and walk to the coast.

Stop 2 - 2 Black Point

Outcrop close to the structural base of the Devereaux ophiolitic fragment shows an example of highly strained gabbroic gneisses, which have been folded into shallowly east plunging F2 folds. The shear zones in the gneisses formed under amphibolite facies conditions, conditions which are not found in any other Ordovician rock type in the Elmtree-Belledune Inlier. These presently shallow dipping shearzones can be traced northwards into stratigraphically higher parts of the gabbro where they have steeper dips. These shear zones formed either during spreading related deformation or during transform faulting. An age difference of 2 to 4 My between the syntectonic plagiogranite dikes and the gabbros (see below) supports a transform faulting environment.

3.1 km Park car on right side of the road and walk along the beach northwards to first outcrop.

Stop 2 - 3 Pointe Verte Melange (Fig.9).

A sequence of wacke/shale rhythmites, similar to those of stop 1, black shale, chert and alkali pillow basalt of the Pointe Verte Formation are strongly deformed during F1. The F1 related deformation caused intense transposition and locally mixing of the various rock types, giving rise for instance to large rafts of chert in a shale matrix. Hence its interpretation as a melange by Rast and Stringer (1980). The F1 folds typically have no associated axial plane cleavage and are recumbent in the hinges of large shallowly eastwards plunging F2 folds. F2 refolding of the overturned limbs of F1 folds produced downwards facing structures. The F2 structures are overprinted by the NE trending F3 folds and cleavages.

4.8 km Proceed northwards for approximately 1.7 km and turn left onto to Rue de la Gare (log 4.8).

5.6 km 2 - 4 Green Point Station. Turn left just before the railway crossing and park car. Walk south to large outcrop along the railway.

Strongly sheared gabbroic gneisses (like in stop 2) of the Devereaux ophiolitic fragment are intruded by diabase and plagiogranite dikes. Zircons extracted from a low strain pegmatitic gabbro pod yielded a concordant U-Pb zircon age of 463.9 +/- 1.0 Ma. (Sullivan et al., 1989). The sheared gabbro is cut by plagiogranite dikes, which are also strained by these steeply dipping high grade shear zones, albeit weaker. The plagiogranite dike has a concordant U-Pb zircon age of 459.6 +/- 1.0 Ma. (Sullivan et al., 1989) and probably intruded off-axis in a transform fault setting.

10.5 km Exit 333 Overpass on route 11

Stop 2 - 5

This roadcut shows the gradual contact between the pillowed alkali basalt and pyroclastic rocks of the Madran Member

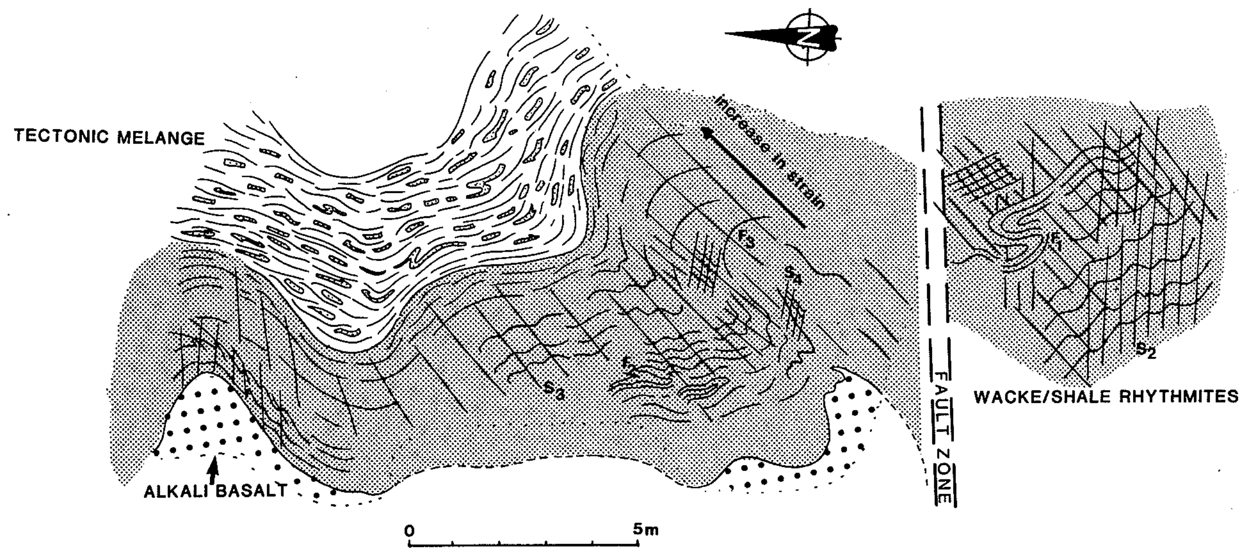

Figure 9. Sketch of small scale structures in the Pointe Verte Formation on approaching the melange in Pointe Verte.

with the wackes and shales of the Quitard Brook Member. Together these two members define the Pointe Verte Formation. The pillows in the northern part of the roadcut face upwards instead of sideways and suggest that the Madran Member has a relatively shallow dip in this part of the area. The alkali basalts have a relatively high chromium content (> 200 ppm) which makes them distinct from any of the other alkali basalt suites in the area. Interpillow limestone yielded a conodont fauna indicative of the P. anserinus zone, which corresponds with a late Llandeilian age (Nowlan, 1983). Higher up in the section these alkali basalts are interbedded with black shale, which yielded graptolites of the Llandeilian-lower Caradocian N. gracilis zone (Riva in Fyffe, 1986).

0 km Proceed northwards on route 11. Restart road log. Approx. 7.0 km, approx. 550 m before the railway crossing, turn right on a small road, which should be followed for approx. 2 km. Then turn right at the first bifurcation in the road and proceed southwards upon reaching a rock quarry filled with bulrushes.

Stop 2-6 Bulrush quarry

The quarry contains a gabbro sill in basalt that is interbedded with wackes and red/green shales of the Devereaux Formation. These basalts have compositions intermediate between MORB and IAT, i.e they are depleted in high field strength elements with respect to N-MORB. Such basalts are typical of back-arc basins (Saunders and Tarney,1984).

Return to Route 11 and proceed northwards for approx. 350 m. and turn left onto a small sandy road. Proceed for approx. 1.6 km and park car at a small clearing. Walk the rest of the road, the last part parallel to the power line, to a big clearing.

Stop 2 - 7 Turgeon

The Turgeon deposit is a small pyritic Zn-Cu deposit, comprising several massive lenses, pods, veins and disseminated sulphide bodies. They are hosted by tholeiitic basaltic rocks interbedded with jasper of the Devereaux Formation. Alteration is locally intense and includes silicification of the basalts.

Return to Route 11 and proceed southwards.

20.5 km exit to Beresford and proceed westwards for 300 m, turn left and follow road, which ends up in a cul de sac for approximately 1 km upon reaching bridge.

Stop 2 - 8 Small bridge across Millstream River in Beresford

Thick bedded lithic and feldspathic wackes of the Middle Ordovician Millstream Formation are rhythmically interbedded with greyish green shales. Grading, cross bedding, rip-up clasts and lamination can be seen in sandy layers in clean washed outcrops during low water levels, suggesting that these rocks represent turbidite deposits.

The Millstream Formation is lithologically very similar to the Pointe Verte Formation in the Elmtree-Belledune Inlier or the sediments that overlie the Devereaux Formation.

Return to major road and proceed westward for approx. 1 km upon reaching intersection with Route 315. Turn left and cross the bridge over the Millstream River in Robertville. Park car in cul de sac on right side of the road. Walk down to the river at the location of the falls behind second house from the road. (Please ask permission).

Stop 2 - 9

Small body of tholeiitic basalt of the Sormany Formation intercalated with the wackes of the Millstream Formation. The basalt is generally aphyric and consists mainly of a fine groundmass of fans of diverging plagioclase (albite) needles. Some vague indications of pillow structures are locally preserved. A narrow fault zone with unknown sense of displacement seems to be located in the river. Despite the presence of a potential fault zone, the isolated nature of this basalt body suggest that the Llandeilian basalts of the Sormany Formation are interlayered with the wackes of the Millstream Formation. The Millstream wackes are thus also, at least in part, Middle Ordovician in age.

Return to Route 11. Restart roadlog and proceed southwards for 4.5 km.

Stop 2 - 10 Rock cut on Bathurst bypass and adjacent quarry

Rock cut and quarry contain pillow basalt, which are interbedded with red shale and chert, pyroclastic breccias and a few diabase sills in the rock cut along the bypass. Basalts and sills are locally vesicular or amygdaloidal

These basalts are strongly alkaline and chemically very distinctive and referred to as the Beresford alkali basalt suite. These basalts, which are interbedded with the sediments of the Boucher Brook Formation, contain differentiates to trachyandesite, trachyte and comendite. An interbedded trachyte yielded a U-Pb zircon date of 457+/-1Ma..Overlying black shale contains an early Caradocian graptolite fauna, probably of the N. gracilis zone (J.Riva, pers. comm.). Both ages suggest that the Beresford alkali basalt suite is late Llandeilian or early Caradocian in age. The pillows in the rock quarry young towards the north but the sedimentary structures indicate that the rocks have a dominantly southwards facing direction in the rock cut along the bypass (Fig. 10). This change in facing direction confirms the presence of an anticline (cf. Skinner, 1956) but is contrary to the interpretations of Rast and Stringer, 1980. Younging indicators are best preserved at the northern end of the outcrop. These include large flames of red mud or silt in the pillow basalt and grading and channelling in the interlayerd red shales and silts. As a consequence the late Caradocian (D.clingani) black shale (Riva and Malo, 1988) that bound these basalts to the south must be in tectonic contact with the latter. Van Staal et al. (1988b) interpreted this fault as a thrust (Fig. 11), which is located in the narrow zone occupied by the strongly deformed black shale (Stop 11).

Proceed southwards 2.0 km Park car on the right side of the road next to a sand pit. Walk through the sand pit along the east trending path to the Tetagouche River. Walk northwards along the river for approx. 300 m. to the bend in the river.

Stop 2 - 11

Black shale of the Boucher Brook Formation is well exposed along the banks of the Tetagouche River for a length of

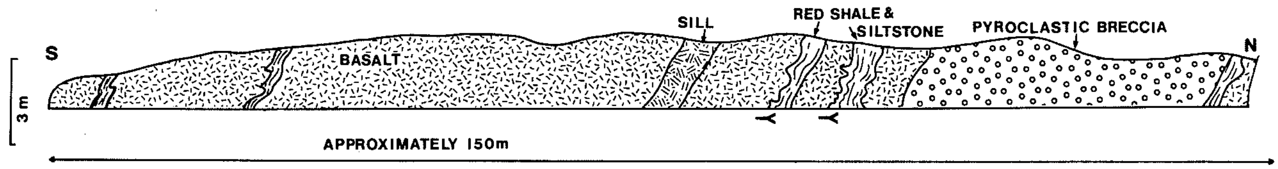

Figure 10. Sketch of the geological relationships in the roadcut on Route 11 just north of Peters river north of Bathurst.

ca. 400 meters going upstream from the railway bridge. The outcrop pattern of the black shale and basalt suggest the presence of a tight z-shaped fold (Fig.11). The black shale yielded abundant graptolites at two localities (van Staal et al., 1988), which indicate a late Caradocian age (D. clingani zone; Riva and Malo, 1988). The basal contact of the black shale with a polymictic conglomerate is highly sheared and is probably faulted. The upper contact with red shale is sharp and devoid of any brittle deformation features. However, these rocks contain a very well developed bedding parallel cleavage, which is thought to have formed by bedding-parallel shear (see above). This boundary between the black and red shales is probably the site of the thrust contact. The strong deformation concentrated along the basal contact of the black shale unit may be due to footwall collapse. If correct the black shale unit is a tectonic horst.

The black shale layer is bounded to the south by respectively a rudaceous to arenaceous limestone unit, and a sequence of thin bedded calcareous siltstones, mudstones and cross bedded limestone.

Together these rocks are combined into the Vallee Lourdes Formation (van Staal et al, 1988b). Large granitoid pebbles in the rudaceous limestone yielded Grenvillian age U-Pb zircon ages (R. Sullivan, pers. comm.) and suggest that Grenvillian age basement was exposed in close proximity during their time of deposition or the pebbles are basement fragments brought up during explosive volcanic eruptions. However the Vallee Lourdes Formation contains middle/late Arenigian Celtic brachiopod faunas (Neuman, 1984) and Arenigian to early Llanvirnian North Atlantic conodont faunas (Nowlan, 1981) indicating significant separation from the North American continent.

The Vallee Lourdes Formation is underlain by the Miramichi Group sediments, comprising quartzite and quartz wacke/shale rhythmites.

The contact between the Vallee Lourdes Formation and the Miramichi Group is not exposed in the Tetagouche River in Bathurst but 10 km. westwards along strike at little Falls. This contact can be shown here to be a disconformity (Fyffe and Noble, 1985).

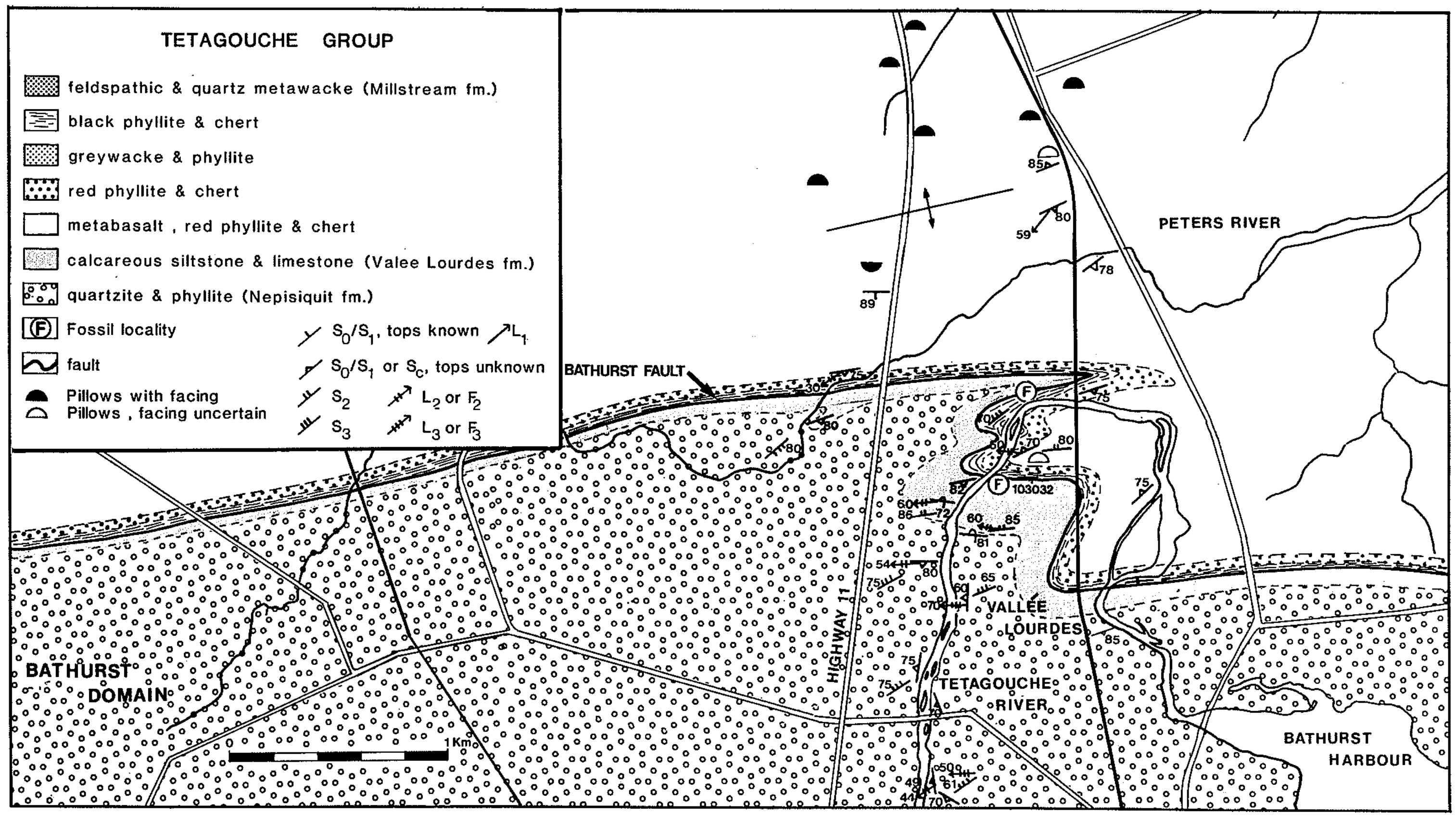

Figure 11. Geology in the vicinity of the town of Bathurst.

ROAD LOG FOR EAST-WEST TRANSECT THROUGH THE TETAGOUCHE GROUP AND PARTS OF FOURNIER GROUP IN THE NORTHERN MIRAMICHI HIGHLANDS. ROAD TOUR NO. 3.

0 km Exit 310 on Bathurst bypass. Proceed westwards on Vanier boulevard past the Bathurst airport where it turns into road to resources (Route 180).

9.8 km Turn right into Tetagouche Falls provincial picnic site.

Stop 3 - 1

The viewing area is underlain by pyroclastic and tuffites/proximal epiclastic rocks of the Nepisiquit Falls Formation, which is the principal unit that underlies most of the Brunswick type massive sulphide deposits. These rocks are here closely associated with- and partly interbedded with a multicoloured (red, green and black) manganiferous shale, which seems to occupy the same stratigraphic position as the Brunswick iron formation unit. The rocks are folded into a large Z-shaped, steeply plunging F2 fold with its axial plane trending approximately parallel to the river (Fig. 2).

24.3 km At the intersection with the Arsenault road towards the south and limestone quarry road to the north. Turn right onto quarry road and immediately turn right again onto small gravel road. Proceed on this road for 500 meters.

Stop 3 - 2

Exposures show tholeiitic pillowed oceanic floor basalts of the Fournier Group. Progressive flattening of the pillows shows that the lavas become progressively more strained in a southward direction towards the basal thrust contact with the Tetagouche Group, which is not exposed here.

Return to road to resources and continue westwards 26.4 km

Stop 3 - 3

Roadcut shows mildly strained pillowed ocean floor basalts of the Fournier Group. Note epidotized pillow selvages and contrast in strain with respect to the previous stop.

37 km Continue westwards.
Stop 3 - 4

Roadcut with small quarry showing contact of highly strained phyllonitic mafic rocks with quartz and feldspar phyric rhyolite.

50 km Turn right onto small gravel road and proceed northwestwards

51.3 km
Stop 3 - 5

Exposure shows highly strained blueschists with a mm.-scale S1 differentiated layering consisting of epidote- and sodic amphibole (glaucophane/crossite)-rich layers. Note the complex deformation that superceded S1, comprising F2, F4 and F5 folds.

0 km Return to Route 180 restart log and continue westwards.

1.8 km
Stop 3 - 6

Roadcut showing the tectonic contact between rhyolite and maroon shale and a bit of mafic phyllonite. The mafic phyllonite can be walked out into blueschists. The rhyolite is transformed into a sericitic phyllonite by the high deformation but can be walked out into a feldspar phyric rhyolite (stop 8).

4.2 km
Stop 3 - 7 K-feldspar phyric rhyolite.

Deformation is very mild and hi-lites the strain contrast with the previous stop. Greenish coloured phyllosilicates are light greenish muscovite (probably phengitic). K-feldspar phenocrysts are generally idiomorphic and show baveno as well carlsbad twins. They are in part altered to chess board albite. This rhyolite body forms the hangingwall to the Caribou massive sulphide deposit with which it is interpreted to be in tectonic contact.

7.0 km
Stop 3 - 8

Caradocian black shale of the Boucher Brook Formation, which represent the youngest rock type known in the Tetagouche Group. South of the road, close to the Camel Back mountain these sediments overlie or are interbedded with lower to middle Caradocian limestone lenses (Nowlan, 1981).

8.0 km
Stop 3 - 9

Roadcut on right side of the road is close to the tectonic contact between the shales of the Boucher Brook Formation and the structurally overlying Camel Back alkali basalt suite. The latter is older than the middle Caradocian limestone; hence old overlies young. The structural contact is marked by maroon and red shaly phyllonites. These alkali basalts generally contain sodic amphiboles, whereas chemically identical basalts south of this contact contain typical greenschist facies assemblages. This indicates that this tectonic contact also marks a sudden jump in metamorphic grade with high pressure rocks overlying lower pressure rocks.

9.3 km
Stop 3 - 10

Roadcut in phyllonitic basalts containing at least locally sodic blue amphibole. These rocks mark the tectonic contact between two chemically different alkali basalt bodies, each incorporated into the blueschist belt. These bodies consist of chromium-poor (Cr < 30 ppm) Camel Back alkali basalts to the southeast and the Eighteen Mile Brook alkali basalts, characterised by intermediate chromium values (200<Cr>30 ppm) to the northwest.

Stop 3-11 Continue from here to the Murray Brook precious metal gossan deposit.

The Murray Brook Precious Metal Gossan Deposit, Bathurst Camp, New Brunswick.

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The Murray Brook precious metal gossan deposit represents the largest (1.9 Mt.; 1.50 gm/tonne Au, 65.5 gm/tonne Ag) of a number of supergene enrichment zones developed over the polymetallic massive sulphide deposits of the Bathurst Camp, New Brunswick (see Fig. 2 of previous section for location). A relatively long period of weathering in this area has resulted in the formation of a thick (up to 70 m.) blanket of gossan over the primary mineral zone at Murray Brook (Fig. 1). Nine main rock units have been identified, namely: Hangingwall Rocks, Barren Gossan, Fertile Gossan, Gossanous Breccia, Quartz Sand, Pyrite Sand, Altered Massive Sulphide, Unaltered Massive Sulphide and Footwall Rocks.

The Barren Gossan unit, identified by preservation of structural/ stratigraphic features and low Au and Ag content, consists mainly of oxidized footwall rocks (deposit recumbently folded) containing disseminated sulphides. The Fertile Gossan constitutes the main economic body of mineralization and consists of both celluralr boxwork and pseudomorhic replacement textural units comprising goethite, quartz (euhedral), secondary silica, plumbojarosite, jarosite, beudantite and cassiterite. Gold- and silver-bearing minerals have not been identified, indicating that both are intimately sequestered in the iron oxides. Gossanous Breccia is considered to be the weathered equivalent of two fault zones which transect the deposit. Quartz Sand, comprised mainly of euhedral quartz (80-90%) and goethite, occurs as lensoid bodies (1-2 m thick) in the interior of the Fertile Gossan; it is considered to represent the weathered equivalent of the hangingwall quartz carbonate unit. Thin (less than 2 m), highly friable, Pyrite Band forms a transition unit between the Fertile Gossan and massive sulphide body. This unit displays no base or precious metal secondary enrichment indicating the abscence of a typical 'fluctuating water table' paleohydrologic regime during gossan formation. For a short distance below this transition unit (1-5m.) the massive sulphide body displays alteration consisting mainly of conversion of chalcopyrite to covellite (without metal enrichment) and dissolution of carbonates. The unaltered massive sulphide body comprises a number of discrete Cu-pyrite, Pb-Zn-pyrite and massive pyrite zones.

On a mass balanced, isovolumetric basis, the Fertile Gossan is enriched in SiO_2, K, Pb, As, Sb, Bi, Hg, Se, Ba, Mo, Cr, Au and Ag; and, depleted in S, Fe, Cu, Zn, Cd, Ca, Mg, Na, P, Mn and Al compared to the unaltered massive sulphide.

Constraints on the period of weathering are at present difficult to ascertain precisely. It would appear from available data that uplift and weathering of the Tetagouche Group has proceeded since middle Silurian. Given local uplift rates and average exhumation values the Murray Brook deposit and other gossan capped deposits in the area probably did not reach the supergene environment until middle to late Cretaceous. Preserved laterite profiles in the area indicate a humid semi-tropical environment during part of the Tertiary.

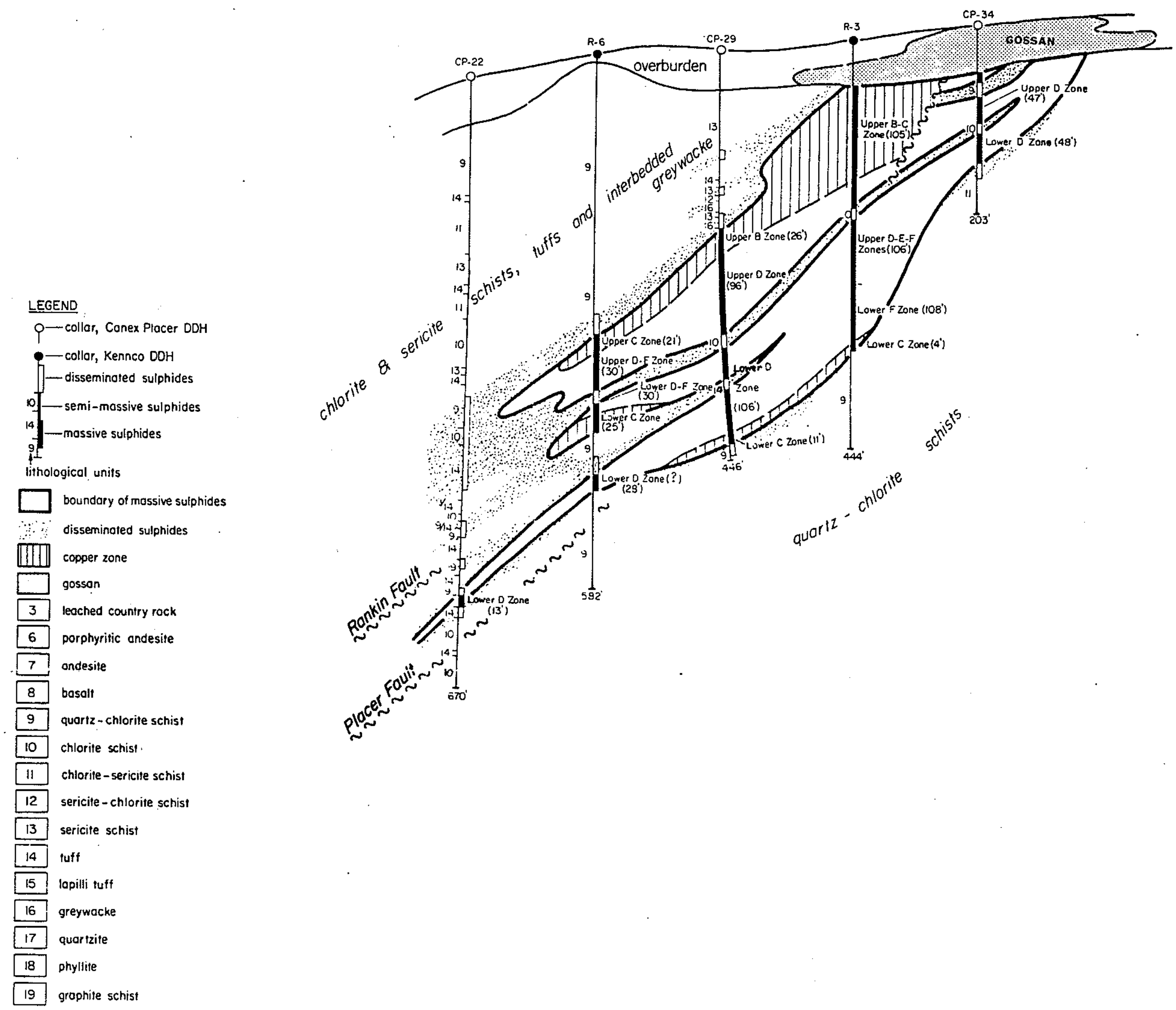

Figure 1. Cross Section (5300E) of Murray Brook massive sulphide deposit (courtesy Kennco Explorations Canada Ltd.).

THE BRUNSWICK NO. 12 AND NO. 6 MINES, BRUNSWICK MINING AND SMELTING CORPORATION LIMITED.

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INTRODUCTION

The Brunswick No. 12 and No. 6 mines are 10 km apart and 27 km southwest of the city of Bathurst, N.B. (Fig. 2 previous section). They occur at or near the same stratigraphic horizon and form part of the Nepisiguit Falls Formation of the Tetagouche Group.

A few diamond-drillholes were put down in Brunswick No. 6 deposit in 1907 during an investigation of the closely associated iron ore (iron formation) similar to that mined in the past at the Austin Brook mine, 0.8 km to the south. However, the No. 6 sulphide deposit was not recognised as such until late in 1952. At this time, interest in the sulphur content of the Austin Brook deposit led to renewed exploration of the area, including a vertical loop electromagnetic survey. Subsequent drilling of strong anomalies led to the discovery of the No. 6 sulphide body, although the first eleven holes of the 1952 drilling campaign were put down in the Austin Brook deposit. Geological Survey of Canada airborne magnetic maps helped delineate the regional trend of lithological units and resulted in extensive staking of magnetic anomalies in 1952-53. The Brunswick No. 12 deposit was discovered in 1953 by drilling a strong electromagnetic anomaly (Davies and Smith, 1966).

Production of ore from the No. 12 deposit started in 1964, followed by No. 6 in 1966 (Table 1). The No. 12 mine has ore reserves of 88,775,000 tonnes grading 8.91% Zn, 3.60% Pb, 0.32% Cu and 101.2 g/t Ag. The No. 6 mine ceased operations in 1983 after producing 12,125,000 tonnes of ore grading 5.43% Zn, 2.16% Pb, 0.39% Cu and 66.5 g/t Ag (Table 1).

STRATIGRAPHY

The stratigraphy of the rocks underlying the No.6 - No. 12 mine area has been discussed above. However, the mine geologists use their own rock nomenclature, which will be explained below.

The stratigraphically lowest rocks in the mine sequence comprise greenish grey quartz wacke rhytmically interbedded with a shaly phyllite of the Miramichi Group. These rocks are generally referred to as the older metasediments (OM)in the mine. Close to the contact with the overlying tuffites and epiclastic rocks of the Nepisiguit Falls Formation the shaly phyllite component increases and becomes more chloritic and graphitic. Small amounts of disseminated pyrite and pyrrhotite are common and small massive sulphide bodies occur locally.

The older metasediments are overlain by quartz-feldspar augen schist (QFAS), quartz eye schist (QES), crystal tuff (CT) and sericite chlorite schist, which respectively refer to quartz- and feldspar phyric rhyolite interlayered with tuffite/epiclastic rocks of the Nepisiguit Falls Formation.

The immediate footwall rocks (FW) to the massive sulphides comprise green, chloritic phyllites with thin beds of wacke, chert and sulphides. Goodfellow (1975) and Juras (1981) interpreted at least part of these rocks as the product of pre-metamorphic alteration related to the formation of the massive sulphides.

The massive sulphide body is divisible into three units or zones:

Table 1. Brunswick Mining and Smelting Pb-Zn-Cu-Ag production for Numbers 6 and 12 Mines, 1964-1989.

	No.12 Mine					No. 6 Mine					No. 6 & No. 12 Mines				
YEAR	Tonnes	%Pb	%Zn	%Cu	g/tAg	Tonnes	%Pb	%Zn	%Cu	g/tAg	Tonnes	%Pb	%Zn	%Cu	g/tAg
1964	877,000	4.06	9.46	0.30	89.00						877,000	4.06	9.46	0.30	89.00
1965	1,504,000	3.96	9.51	0.30	95.00						1,504,000	3.96	9.51	0.30	95.00
1966	1,497,000	3.64	9.26	0.22	76.00	273,000	2.75	6.19	0.35	59.00	1,770,000	3.50	8.79	0.24	73.00
1967	1,514,000	3.47	9.07	0.29	82.00	786,000	2.93	5.96	0.40	61.00	2,300,000	3.29	8.01	0.33	75.00
1968	1,564,000	3.38	8.56	0.27	66.00	892,000	2.47	5.66	0.35	52.00	2,456,000	3.05	7.51	0.30	61.00
1969	1,538,000	3.04	8.05	0.33	73.00	974,000	2.28	5.78	0.35	64.00	2,512,000	2.75	7.17	0.34	70.00
1970	1,343,000	2.93	7.54	0.32	75.00	999,000	2.12	5.86	0.33	63.00	2,342,000	2.58	6.82	0.32	70.00
1971	1,421,000	3.25	8.11	0.30	84.00	768,000	2.11	5.76	0.36	64.00	2,189,000	2.85	7.29	0.32	77.00
1972	1,365,000	3.62	9.10	0.28	96.00	1,593,000	2.05	5.48	0.37	72.00	2,958,000	2.77	7.15	0.33	83.00
1973	1,701,000	3.43	8.58	0.34	96.00	1,317,000	2.02	4.98	0.35	75.00	3,018,000	2.81	7.01	0.34	87.00
1974	1,429,000	3.58	8.00	0.35	89.00	953,000	2.03	4.65	0.41	65.00	2,382,000	2.96	6.66	0.37	79.00
1975	1,982,000	3.54	8.48	0.36	91.00	1,126,000	1.93	4.94	0.50	68.00	3,108,000	2.96	7.20	0.41	83.00
1976	1,473,000	3.53	8.63	0.32	96.00	774,000	1.60	4.41	0.50	68.00	2,247,000	2.87	7.18	0.38	86.00
1977	2,456,000	3.51	8.70	0.31	91.00	678,000	1.70	4.65	0.58	61.00	3,134,000	3.12	7.82	0.37	85.00
1978	2,685,000	3.77	9.31	0.28	97.00	373,000	2.07	5.85	0.39	72.00	3,058,000	3.56	8.89	0.29	94.00
1979	2,722,000	3.71	9.13	0.30	97.00	249,000	2.60	6.78	0.41	78.00	2,971,000	3.62	8.93	0.31	95.00
1980	1,746,000	3.61	8.90	0.31	99.00	102,000	2.75	7.18	0.31	75.00	1,848,000	3.56	8.81	0.31	98.00
1981	3,401,000	3.51	8.75	0.35	98.00	22,000	2.69	7.38	0.14	84.00	3,423,000	3.50	8.74	0.35	98.00
1982	3,535,000	3.64	9.04	0.31	102.00	99,000	3.15	8.62	0.15	94.00	3,634,000	3.63	9.03	0.31	102.00
1983	3,264,000	3.56	9.00	0.30	99.00	147,000	2.57	6.90	0.52	86.00	3,411,000	3.52	8.91	0.31	98.00
1984	3,560,000	3.57	8.91	0.32	100.00						3,560,000	3.57	8.91	0.32	100.00
1985	3,312,000	3.52	8.77	0.32	101.00						3,312,000	3.52	8.77	0.32	101.00
1986	3,409,000	3.54	8.68	0.34	101.00						3,409,000	3.54	8.68	0.34	101.00
1987	3,447,000	3.56	8.94	0.37	104.00						3,447,000	3.56	8.94	0.37	104.00
1988	3,494,000	3.58	8.85	0.36	105.00						3,494,000	3.58	8.85	0.36	105.00
1989	2,983,000	3.56	8.93	0.39	110.00						2,983,000	3.56	8.93	0.39	110.00
Total	59,222,000	3.55	8.82	0.31	97.00	12,125,000	2.16	5.43	0.39	67.00	71,347,000	3.31	8.23	0.33	92.00

W.Luff

1) a massive pyrite zone, containing minor amounts of sphalerite and galena, and minor to large amounts of chalcopyrite, magnetite and pyrrhotite (SPP or SPPC).

2) massive banded pyrite-sphalerite-galena with minor chalcopyrite and pyrrhotite (SO); the latter two minerals becoming more abundant below the 850 level in the No. 12 deposit.

3) massive pyrite comprising very fine grained pyrite, with minor sphalerite, galena and chalcopyrite (SP).

The sulphides are overlain by iron formation (IF), which can be dominated by oxide, silicate or carbonate minerals. Contacts between the various types of iron formation and the massive sulphides are generally gradational. All sorts of mixtures occur and clearly suggest that the sulphides and iron formation form an integral unit of metalliferous sediment.

The immediate hanging wall rocks (HW) consist of grey chloritic and sericitic phyllitic sediments. At No. 6 mine the hanging wall rocks include aphyric or feldspar phyric rhyolite and lapilli tuff, which form part of the Flat Landing Brook Formation.

The No. 6 and No. 12 mine rock sequence is capped by pillowed alkali basalts (B) of the Boucher Brook Formation. The contact between the basalts and the underlying rock units is generally marked by a red or green Fe/Mn-rich shaly or silty phyllite.

The hanging wall rocks of the No. 6 mine are intruded by a southwesterly plunging body of tholeiitic gabbro. A similar gabbroic body has been intersected during underground drilling to the north of the No. 12 mine area. A quartz and feldspar phyric porphyry dike intruded the No. 12 ore body. Mesostructures and greenschist facies mineral assemblages indicate that these intrusive bodies predate regional deformation and metamorphism.

STRUCTURE AND DISTRIBUTION OF THE MASSIVE SULPHIDES

Structural analysis of the Brunswick No. 12 and No. 6 mines and surroundings shows that the structural history and the geometries of the two orebodies are essentially the same as shown in the simplified diagrams in figures 1, 2, 3 and 4. They both occur in large asymmetrical F2 folds, which show a marked variation in plunge but a constant axial plane orientation. On basis of the regional stratigraphy, metal zoning and the stratigraphic position of the iron formation with respect to the sulphides, the F2 folds can be divided into upward and downward facing structures. For instance the large, steeply south plunging Z-shaped fold in the Brunswick No. 12 mine is downward facing. At depth this fold changes plunge by more than 90° and passes through the vertical and then the horizontal to plunge shallowly towards the south such that it becomes upward facing (Fig. 1 and 5). The trace of the F2 fold plunge drawn in a section parallel to the axial plane of the F2 fold therefore defines a large overturned F1 fold (Fig. 5). The large scale geometries of both mines, which define overturned, asymmetrical basins, are therefore interpreted as interference structures between F1 and F2 folds.

Massive sulphides in the Brunswick No. 12 mine occur in four major zones, the Main Zone, the East Zone, the West Zone and the V-2 zone (Fig. 6 and 7) These zones all join below the 850 metre level, where the F2 folds are formed close to a F1 fold hinge, and where the metalliferous rocks can be expected to be thick. The separation of the sulphides on the shallower levels is therefore better explained as a result of the attenuation and faulting that is generally found along the short limbs of the F2 folds rather than interpreting them as separate bodies deposited in second order basins. The attenuation of alternate limbs of almost isoclinal folds resulted in an en echelon pattern of tabular sulphide bodies that correspond to the hinges and long limbs of F2 folds. Numerous intrafolial isoclinal F1 and F2 folds in the sulphide bodies attest to the intense transposition these rocks experienced. The marked boudinage of the porphyry dike in the massive sulphide body (Fig. 6) is another indication of the high strain experienced by the sulphide bodies.

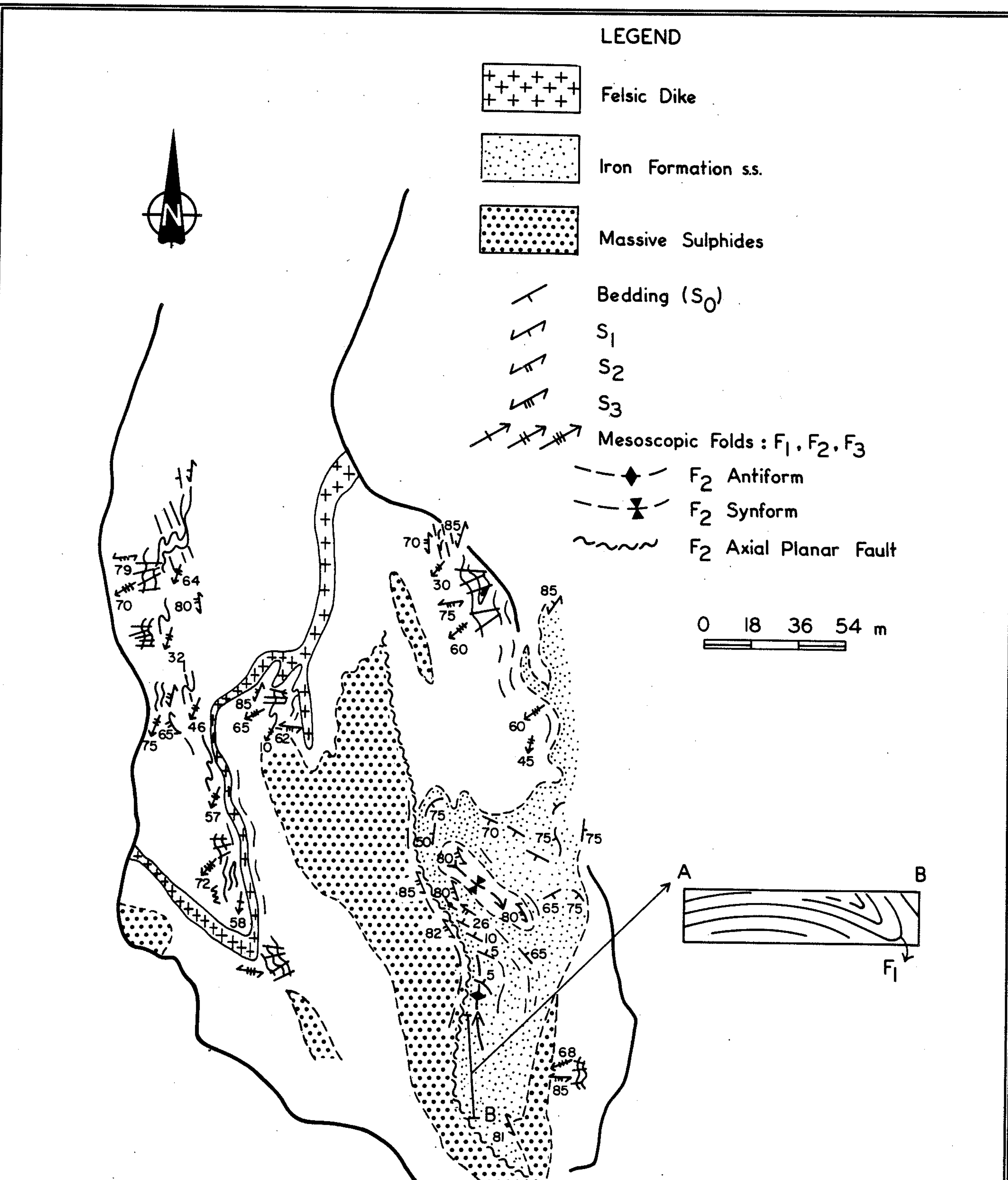

Figure 2. Geology of the Brunswick No.12 open pit. S3 is now changed to S4 (after van Staal and Williams, 1984).

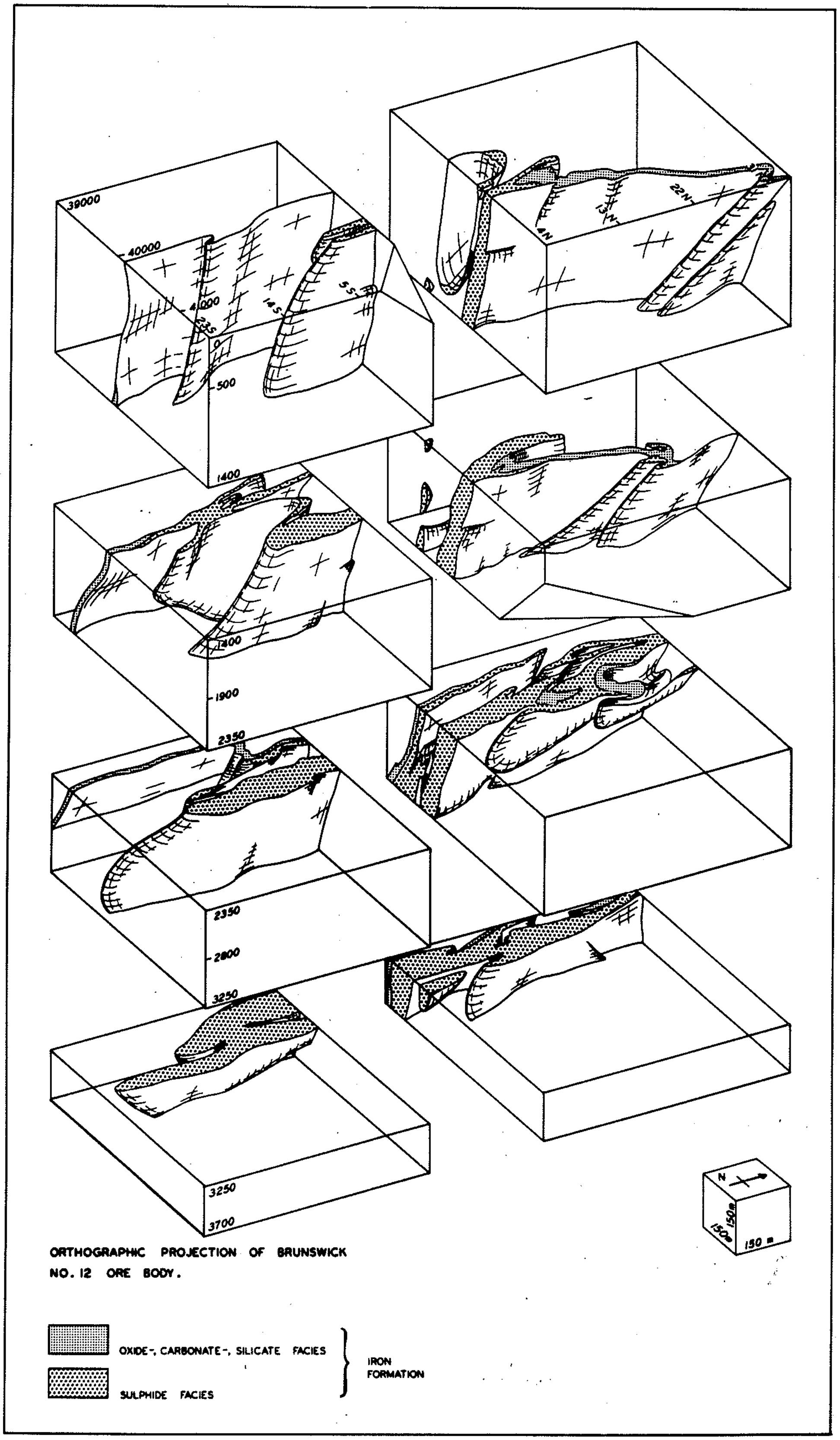

Figure 1. Block diagram of the Brunswick No. 12 orebody (after van Staal and Williams, 1984).

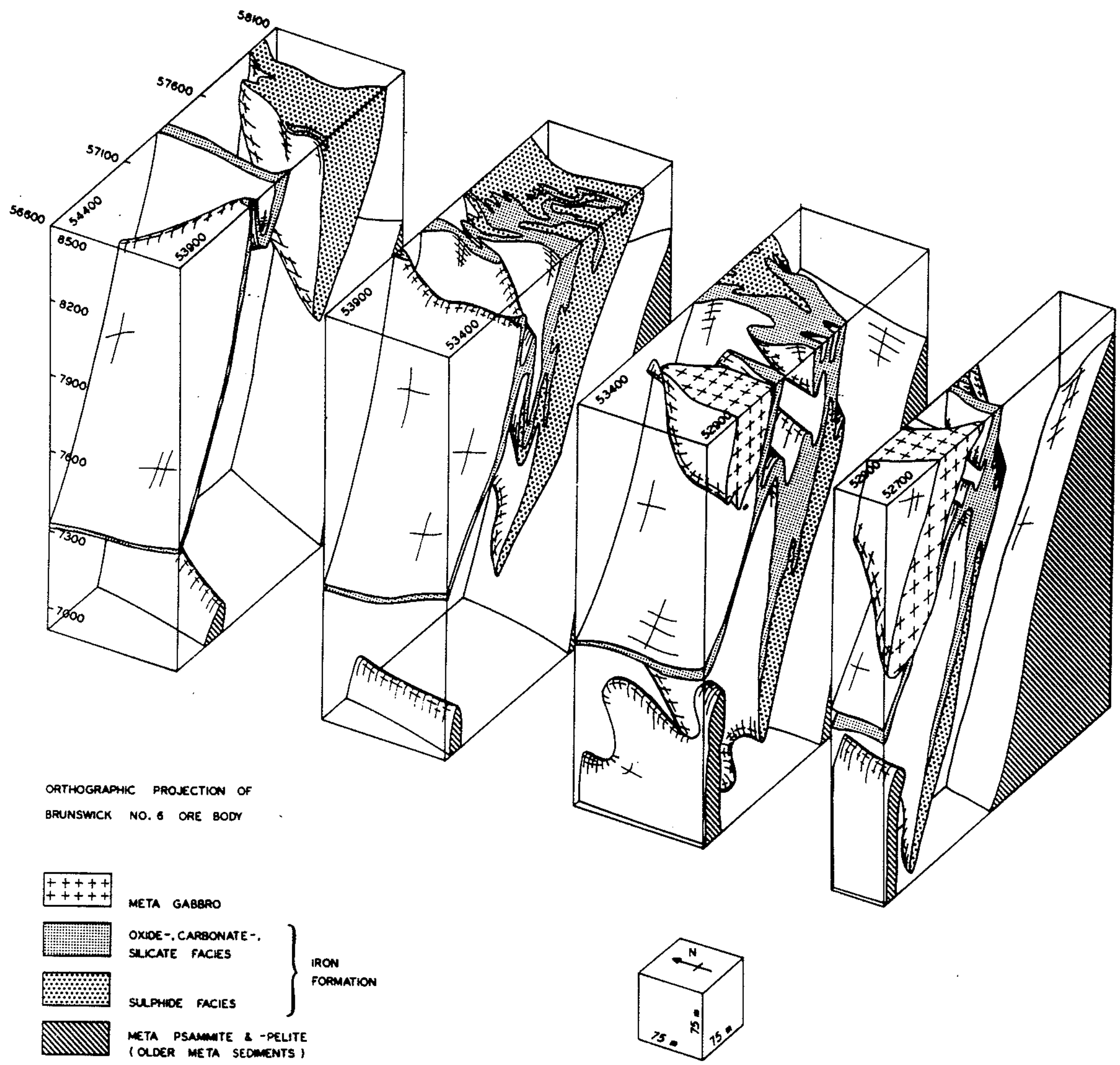

Figure 3. Block diagram of the Brunswick No. 6 orebody (after van Staal and Williams, 1984).

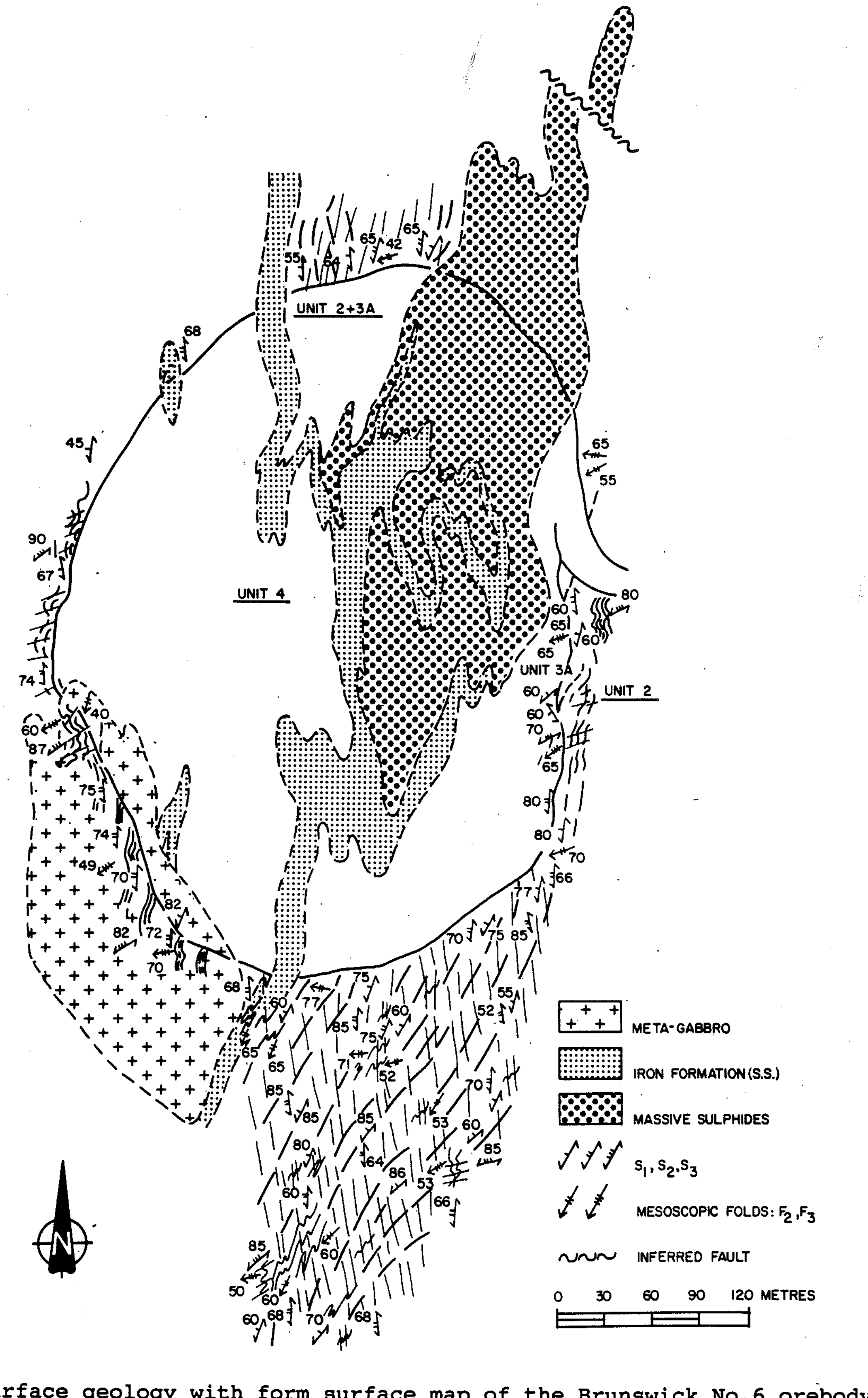

Figure 4. Surface geology with form surface map of the Brunswick No.6 orebody before open pit mining (after Boyle and Davies, 1964 and van Staal and Williams, 1984).

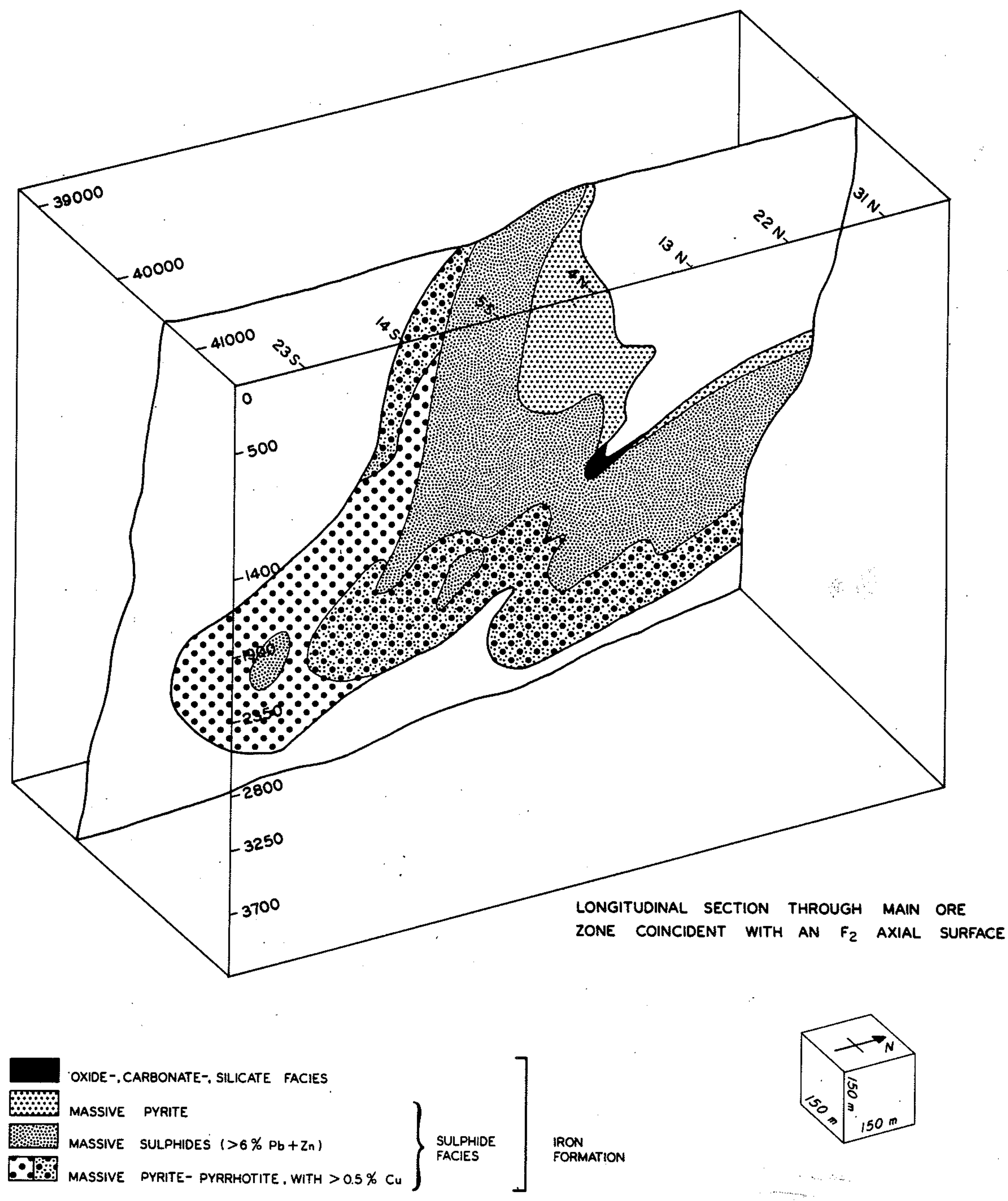

Figure 5. Longitudinal section, Main Zone, Brunswick No.12 orebody parallel to an F2 axial plane (after van Staal and Williams, 1984).

LEGEND

	PORPHYRY DYKE
B	BASIC VOLCANICS - BASALT & BASIC IRON FORMATION
HW	HANGING - WALL METASEDIMENTS & ACID TUFF
IF	IRON FORMATION
SP	MASSIVE PYRITE
SO	MASSIVE SULPHIDES (>6% Pb - Zn)
SPP | SPPC	MASSIVE PYRITE - PYRRHOTITE (>0.5% Cu)
FW	FOOTWALL METASEDIMENTS
CT	CRYSTAL TUFF
M	METASEDIMENTS (CHLORITIC SEDIMENTS)
QES	QUARTZ - EYE SCHIST
OM	OLDER METASEDIMENTS
	F_2 (OR F_1) ANTIFORM
	F_2 (OR F_1) SYNFORM
	CONTACT, APPROXIMATE
	FAULT
23,500 N 12,000 E	MINE COORDINATES

Legend for Figure 6 and 7.

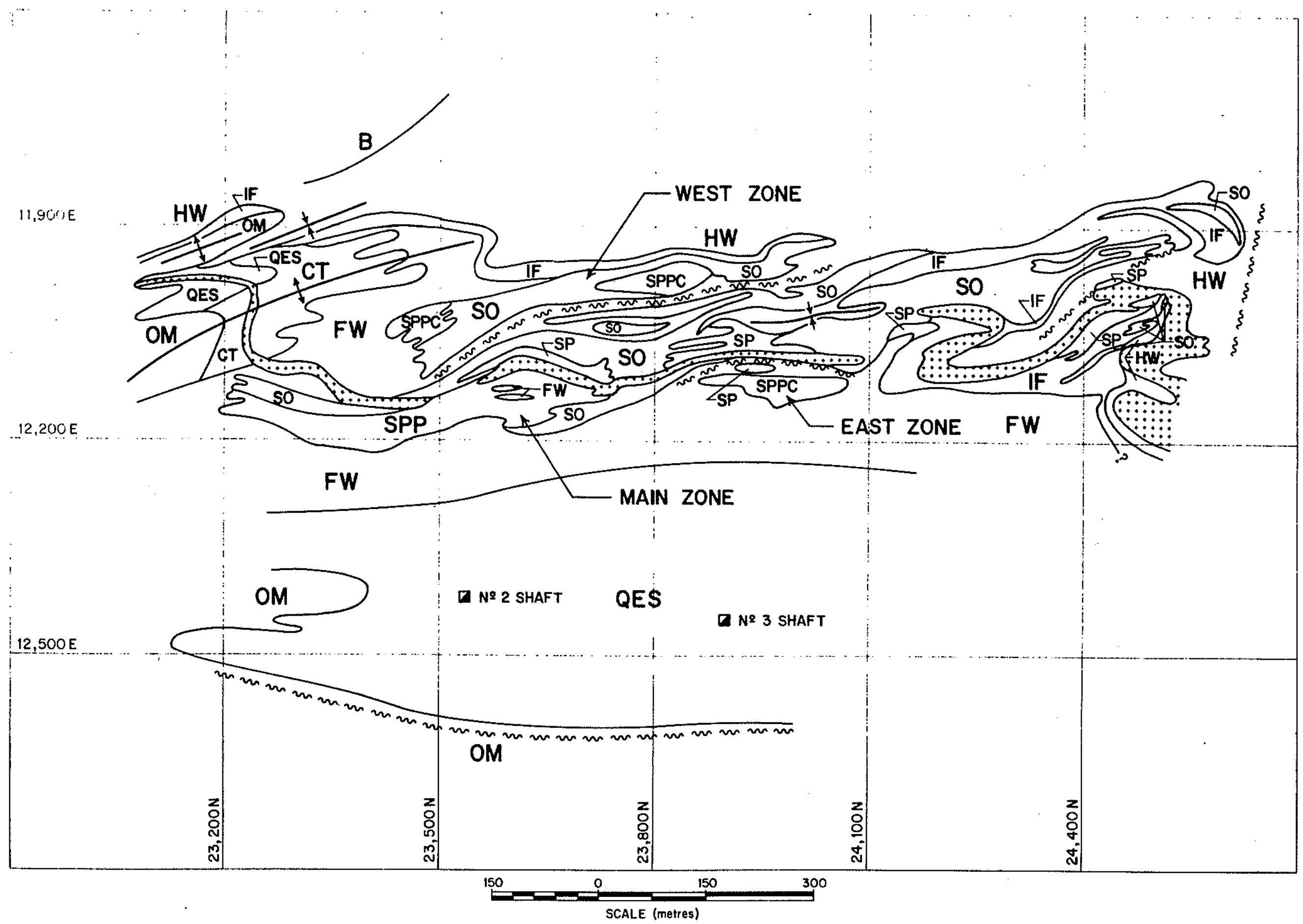

Figure 6. Geological map of the 850 m level of the Brunswick No. 12 orebody.

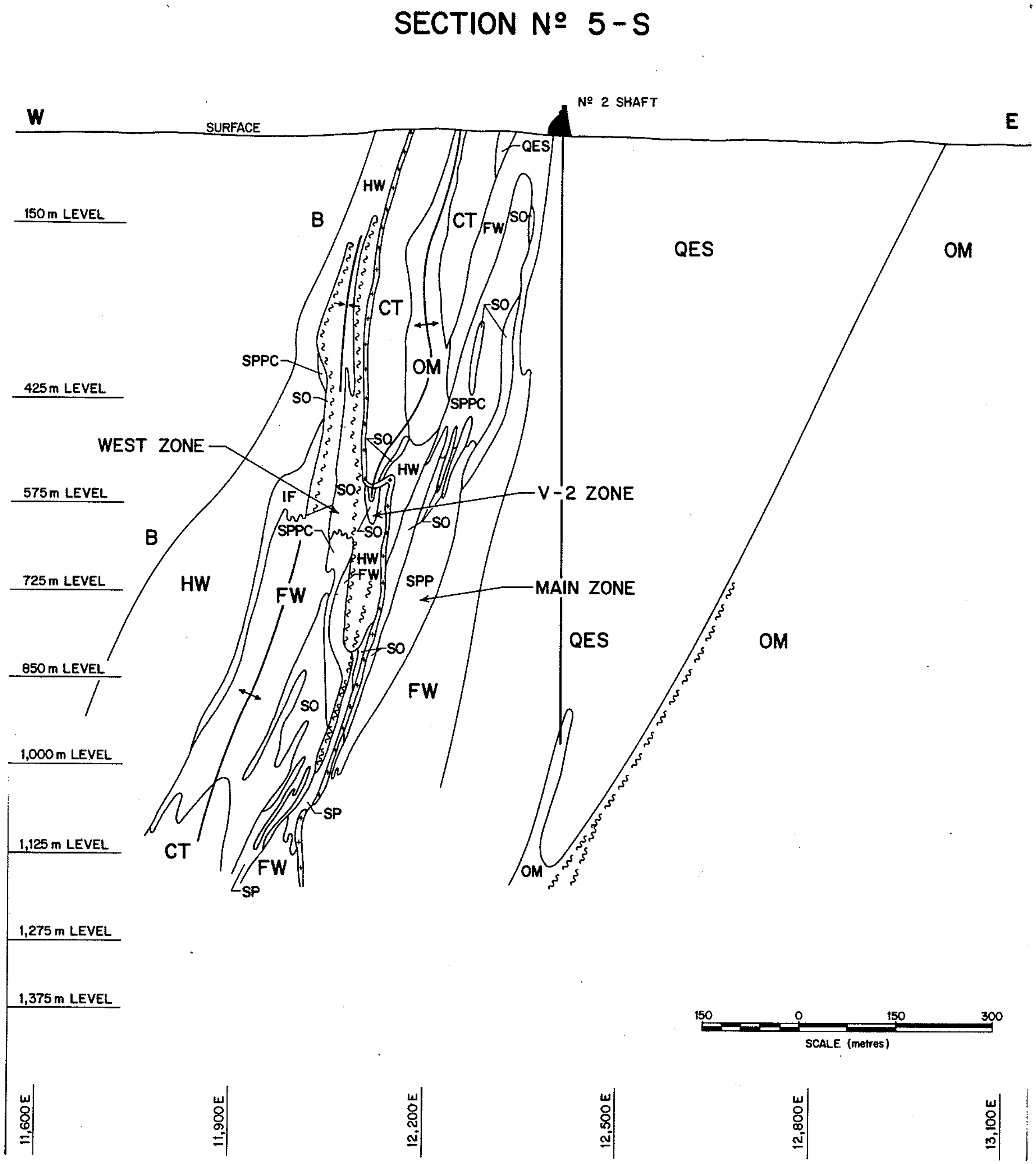

Figure 7. Cross section through the Brunswick No.12 orebody.

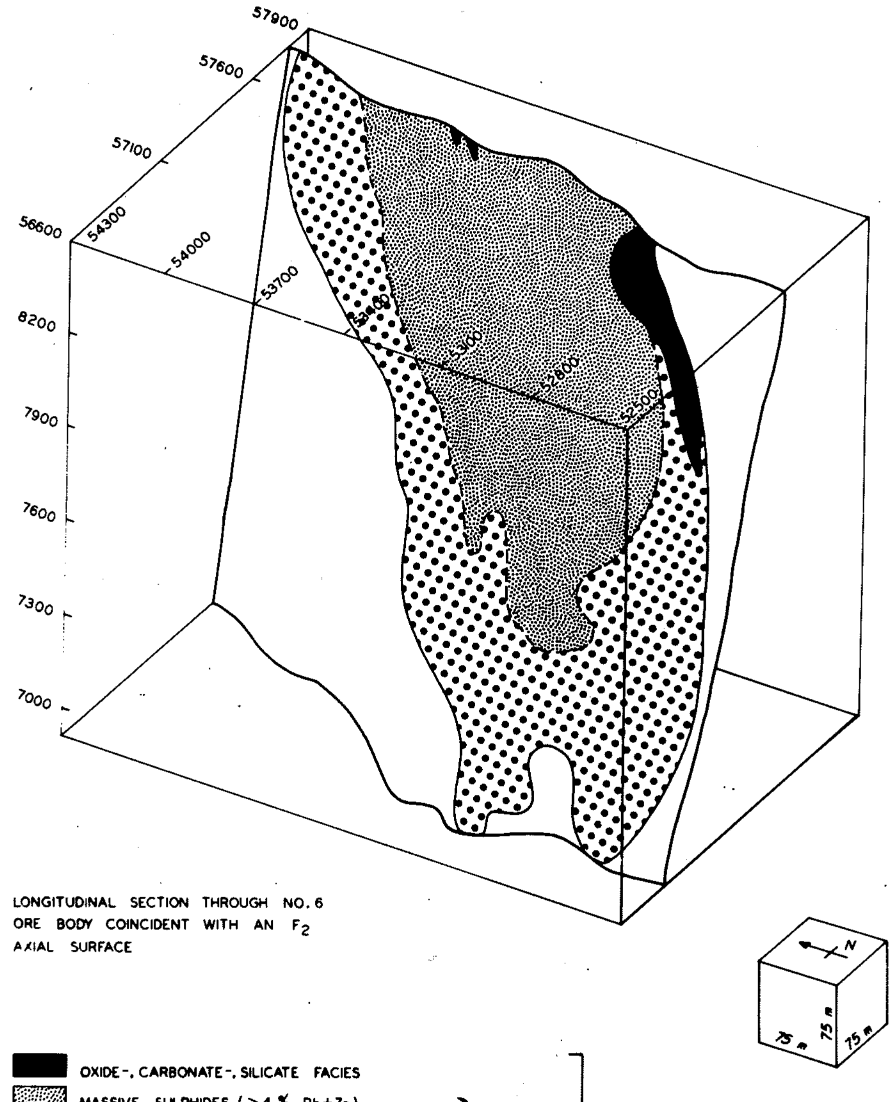

Figure 8. Longitudinal section of the Brunswick No. 6 orebody parallel to an F2 axial plane (after van Staal and Williams, 1984).

ORIGIN OF THE SULPHIDES

Sections parallel to the F2 axial surfaces (Fig. 5 and 8) show that the metal zoning is folded by F1 and probably predates the earliest deformation. All other structural data also indicate that the mineralization, with the exception of some remobilized material, has been affected by the earliest deformation recorded in the country rocks. The structural evidence is thus compatible with a volcanogenic-exhalative origin of the ores. However, proximal features such as a feeder pipe-stringer zone and syngenetic alteration are either missing or obscured by deformation and later alteration. At least part of the cross-cutting sulphide stringers are parallel to the axial surfaces of F1 and F2 folds (Van Staal and Williams, 1984) and therefore cannot be primary. The origin of the folded sulphide stringers, which are generally parallel to layering in the phyllites is hard to determine, since most if not all of the primary relationships betweent the stringers and the footwall sediments have been obscured by the strong deformation concentrated in the incompetent phyllites. It is possible that part of the stringers are related to the ore forming event but this is difficult to prove with the presently available data set.

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THE BURNTHILL GRANITES AND RELATED TUNGSTEN-MOLYBDENUM AND TIN DEPOSITS[1]

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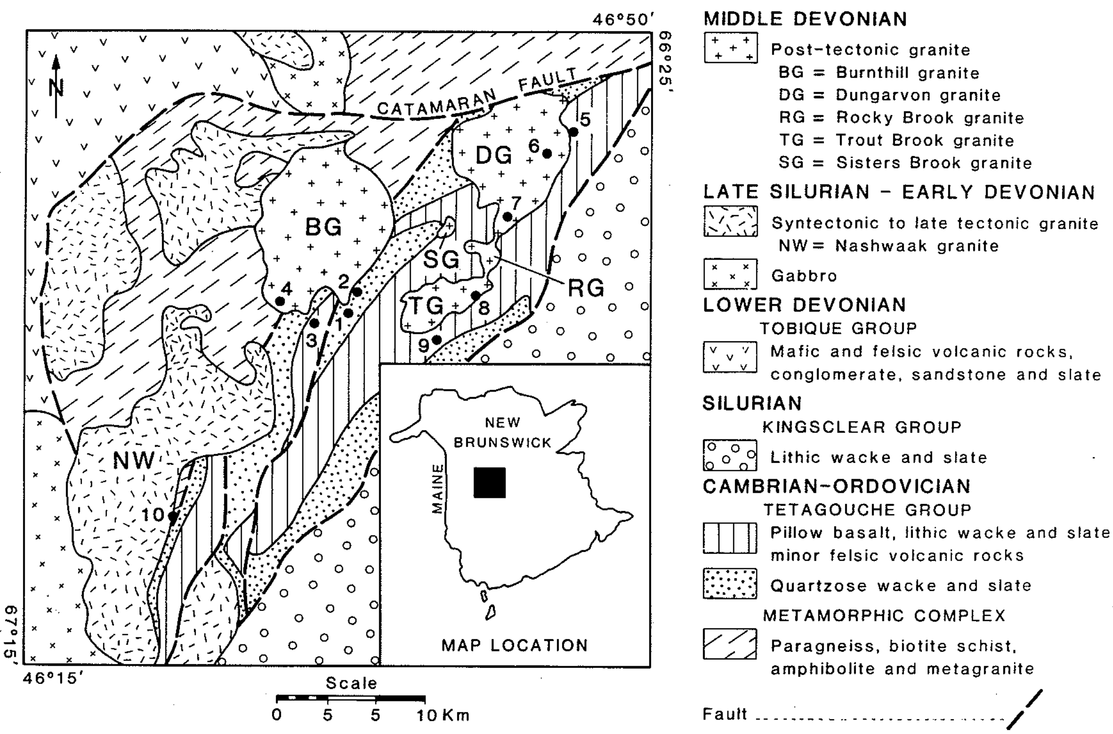
General geology

The Burnthill Granites are a cluster of high-level, multi-phase, high silica, post-tectonic Middle Devonian (Taylor et al., 1987) granite plutons located within the Miramichi highlands of central New Brunswick (Fig. 1). The granites are known as the Burnthill, Dungarvon, Trout Brook, Rocky Brook and Sisters Brook granites. At least one small satellite stock (Tin Hill) is also known. The Dungarvon, Rocky Brook and Trout Brook granites now appear to be connected at depth (Gardiner and Garnett, 1986, 1987a, 1987b) (Fig. 2).

The granites intrude the southwestward extension of the Cambro-Ordovician Tetagouche Group which consists of a thick sequence of quartz wackes and slates with several intercalated pillow basalt horizons and minor felsic volcanic rocks. Calcareous siltstones occur within the wackes. Deformation of these supracrustal rocks occurred during the late Middle Ordovician Taconian orogeny (Fyffe, 1982). Metamorphism related to this deformation ranges from greenschist grade in the east to amphibolite grade in the west.



MINERAL OCCURRENCES

1. Burnthill (W,Sn,Mo,Be,F)
2. Tin Hill (W,Sn,F)
3. McLean Brook South (Sn,Zn, Mo,Cu)
4. McClean Brook North (W)
5. Peaked Mountain (W,Mo)
6. Harris Lake (U,F)
7. Little Dungarvon River (Main Dungarvon and Four Mile Brook) (Sn,F)
8. Trout Brook East (Sn,F)
9. Todd Mountain (Sn,Cu)
10. Sisson Brook (W,Cu,Mo)

Figure 1. Geological setting of granites in the Burnthill Brook area.

[1]This paper was published previously in a Geological Association of Canada Guidebook by Procyshyn et al. (1989) and is reprinted here with minor revisions.

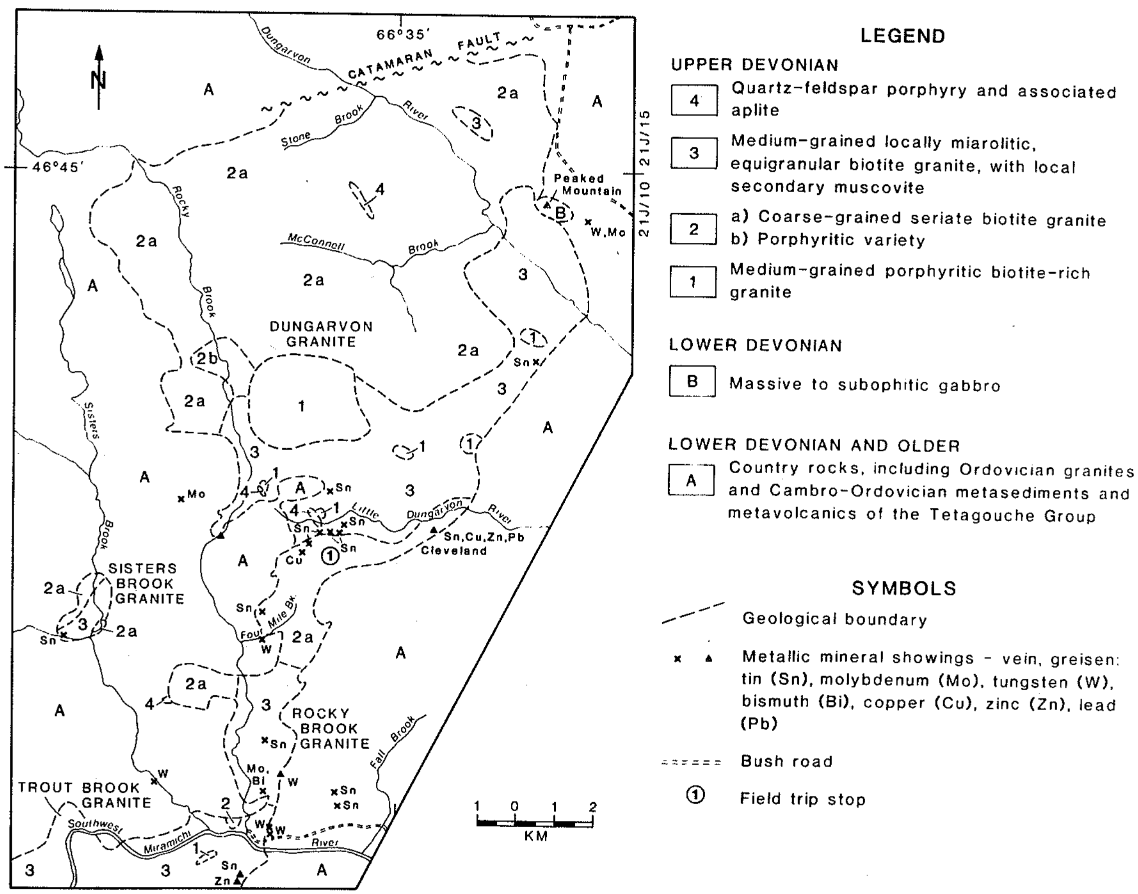


Figure 2. Geological map of the Dungarvon, Rocky Brook, Sisters Brook and Trout Brook (north half) granites.

The Burnthill Granite has also intruded a series of high-grade metamorphic rocks along its northwestern margin. This sequence is comprised of interlayered paragneiss, biotite schist, and amphibolite intruded by concordant plutons of metagranite. The stratigraphic relationship of these rocks to the Tetagouche Group is uncertain (Fyffe and Cormier, 1979).

Accompanying the Acadian orogeny were a series of generally massive, elongated plutons (of which the Nashwaak Granite is one). These granites vary in composition from granite to granodiorite and may exhibit agmatitic margins as is seen near the Sisson Brook W-Mo deposit. This is in contrast to the less voluminous Burnthill granites which are generally subcircular, multi-phase, higher-level undeformed plutons.

The majority of the W-Sn-Mo-U mineral occurrences are spatially associated with the Burnthill granites; they include the Burnthill (W-Mo-Be-F-Cu), Tin Hill (Sn-W-F), McLean Brook North (W), Peaked Mountain (W-Mo), Harris Lake (U), Little Dungarvon (Sn-F), Trout Brook East (Sn-F), and Todd Mountain (Sn-Cu) occurrences. The one exception is the Sisson Brook (W-Mo-Cu) deposit near the eastern margin of the older Nashwaak Granite. A small undated feldspar-biotite porphyry plug is associated with the south mineralized zone of the Sisson Brook deposit and this pluton may be genetically related to the Burnthill granites to the north.

The multi-phase Burnthill granites have been described by MacLellan et al. (1986), MacLellan and Taylor (1989), Taylor et al. (1987) and Gardiner and Garnett (1986, 1987a, 1987b) (Fig. 2 and Fig. 3). The various phases are similar for all the plutons in the group. Unit 1 consists of a medium-grained porphyritic biotite granite characterized by a relatively high biotite content (about 5%), coarse feldspar phenocrysts and, locally, by a higher plagioclase content. Muscovite occurs locally as an accessory mineral and in miarolitic cavities.

Unit 2 is considered the main phase and contains a variety of texturally different granitic rocks that are considered to be comagmatic. The bulk of the unit consists of a medium- to coarse-grained seriate biotite granite that is coarsest to the northwest (at least in the Dungarvon Granite) and at lower elevations. A fine grained marginal phase is locally present along the metasedimentary contact. This unit becomes porphyritic in places with very coarse feldspar phenocrysts occurring in a medium grained groundmass. This unit is generally in gradational contact with Unit 3 over widths of up to several hundreds of metres although crosscutting relationships do occur.

Unit 3 is a fine- to medium-grained, equigranular biotite granite (with muscovite in the Trout Brook Granite), that locally contains quartz phenocrysts, particularly near its contact with Unit 2. Feldspar phenocrysts are scarce. Miarolitic cavities, many of which are open or fluorite-filled, are common. The finer grained portions of this unit locally grade into feldspar $\pm$ biotite pegmatite veins or lenses which are often mineralized with cassiterite, flourite or epidote occurring in miarolitic cavities. This unit has been observed in boulders to cut units 1 and 2.

Unit 4 comprises a number of dyke phases, represented by quartz-feldspar porphyry, aplite, and composite dykes. The composite dykes are of two types: a biotite-banded fine- to medium-grained phase and a quartz-feldspar porphyry that would appear from boulders to have an associated mafic phase. This latter dyke phase may be significantly younger than the plutons they intrude.

The granite phases exhibit systematic decreases in TiO_2, Al_2O_3, MgO, CaO, P_2O_5, Ba, Sr, Zr, and V and increases in SiO_2 from units 1-4 (Fyffe and MacLellan, 1988). The granites are markedly enriched in SiO_2, K_2O, Rb, Sn, F and depleted in CaO, MgO, Sr and B. These features are characteristic of slightly peraluminous anorogenic granites. Sn contents range from 5-15 ppm for unit 1, to 15-25 ppm for unit 2, and to 25-50 ppm for unit 3. Unit 4 tin values are lower, in the range of 5-20 ppm.

Mineral Occurrences

Numerous mineral occurrences are found throughout the area of the Burnthill Granites. They fall into two general categories: endogranitic and exogranitic types. Each type can be subdivided into several subtypes as is shown below:

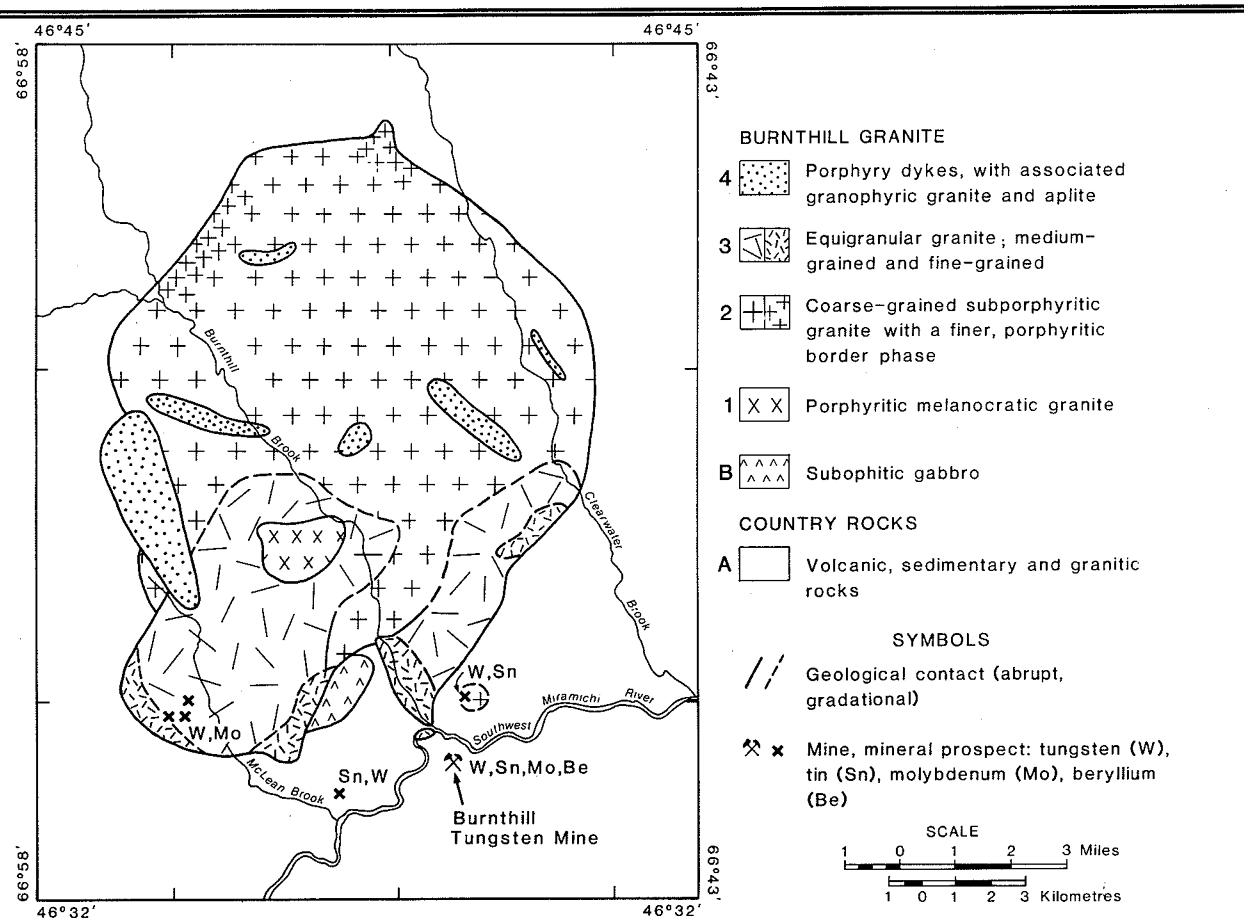


Figure 3. Geological map of the Burnthill Granite pluton (modified from Taylor et al., 1987).

A. ENDOGRANITIC

1. Feldspar (pegmatitic) veins
   a. Magmatic -- 2 feldspars ± biotite
   b. Hydrothermal -- K-feldspar fracture infillings
   c. Hydrothermal -- K-feldspar replacement of granite

2. Quartz-feldspar veins
   a. Feldspar in brecciated remnants of type 1 vein in quartz matrix, or
   b. Hydrothermal feldspar in quartz younger than type 1 feldspars
   c. Minor feldspar + chlorite in quartz

3. Quartz veins (without greisen selvedges)
   a. 2-5 generations of quartz becoming whiter (more opaque) as fluids cool ending with chalcedony-jasper veins
   b. Euhedral quartz as cavity infillings and in aplite veins where quartz growth precedes aplite injection

4. Sulphate-carbonate veins

5. Greisen veins
   a. Quartz veins with greisen selvedges
   b. Lenses, pods and linear zones of quartz and muscovite (i. 90:10 or ii. 10:90)
   c. Greisenized granite (trace to intense)

6. Sulphide disseminations
   a. Associated with grey-green (weak greisen) alteration

of granite
b. Associated with red (hematitic) alteration of granite

7. Sulphide stringers
   a. Cu + Sn sulphides
   b. Mo

8. Breccias
   a. Hydrothermal
   b. Tectonic

B. EXOGRANITIC

9. Scheelite/molybdenite/molybdoscheelite/quartz veins
   a. Sheeted
   b. Stockwork

10. Quartz ± cassiterite ±wolframite ± Cu, Zn, Be, F, Pb
   a. Sheeted
   b. Solitary
   c. Silica-flooded zones

11. Basemetal ± Sn ± W ± F veins

12. Feldspar (pegmatitic) veins

13. Granitic dykes

14. Skarn/hornfels
   a. Calcareous siltstones
   b. Mafic volcanics

Type 1a mineralization (feldspar veins) occurs as miarolitic cavity infillings of cassiterite, epidote, fluorite and/or an earthy material (manganese oxide ?) and as disseminations of cassiterite and pyrite (oxidized) within feldspar (pegmatitic) veins and lenses. These veins generally crosscut granite but are often in gradational contact over 10-20 cm with the host granite suggesting a late magmatic origin for the veins. This type also has high uranium contents (>30 ppm U) and also elevated Nb contents. Feldspar veins may also be tectonized resulting in a finely ground feldspar-cassiterite rock often with later pyrite disseminations. Examples of type 1a mineralization include Little Dungarvon, Harris Brook, and Trout Brook East occurrences.

Type 1b mineralization is characterized by bands of cassiterite and/or disseminations of pyrite within feldspar veins. Feldspar veins may be either pyrite-free or contain abundant pyrite. This bimodality of iron content may also be reflected in the cassiterites which are frequently zoned with black (Fe-poor) cassiterite overgrown by red-brown (Fe-rich) cassiterite. The Four Mile Brook occurrence is an example of this type of mineralization.

Type 1c cassiterite and pyrite (+ chalcopyrite ?) mineralization occurs as disseminations in altered granite (e.g. Tin Hill area).

In Type 2 quartz-feldspar veins, quartz is light grey and nearly transparent and frequently carries cassiterite. Type 2 veins are probably derived from, and grade into, type 1 veins. The feldspars and cassiterites in the quartz are frequently fractured. Cassiterites enclosed by feldspar crystals are generally unfractured. Nb may be as high as 1800 ppm (in a sample assaying 15% Sn) in type 2a veins. Types 2a and 2b contain 10-50% feldspar grading to type 2c which has only 5% feldspar but with increasing chlorite content (to 5%). Types 2a and 2c are represented by the Little Dungarvon occurrence, type 2b is represented by the Rocky Brook occurrence.

Type 3a quartz veins are generally barren of Sn and W but locally contain abundant Mo, Bi and specular hematite. Type 3b veins are frequently open in the centre (or sulphate-infilled, typically anhydrite). Quartz occurs as needles and coarser euhedral crystals and typically is zoned with a clear centre and a thin milky white rind. An example of type 3a is the Rocky Brook occurrence, examples of type 3b are the Little Dungarvon and Rocky Brook occurrences.

The type 4 carbonate-sulphate veins are generally thin (1-2 mm) and coat joint and fracture surfaces. They have been observed to be as much as 2 cm thick in association with quartz veins. Their composition is calcite and anhydrite, although one XRD determination suggests anglesite ($PbSO_4$) is also present. The Little Dungarvon and Rocky Brook) occurrences include examples of the type 4 mineralization.

Types 1, 2b and some 3a veins are associated with red hematite alteration of the granitic wall rocks. Types 2c, 3 and 4 are associated with hematite and illite (green clay) alteration while the milky white quartz and the chalcedonic veins are

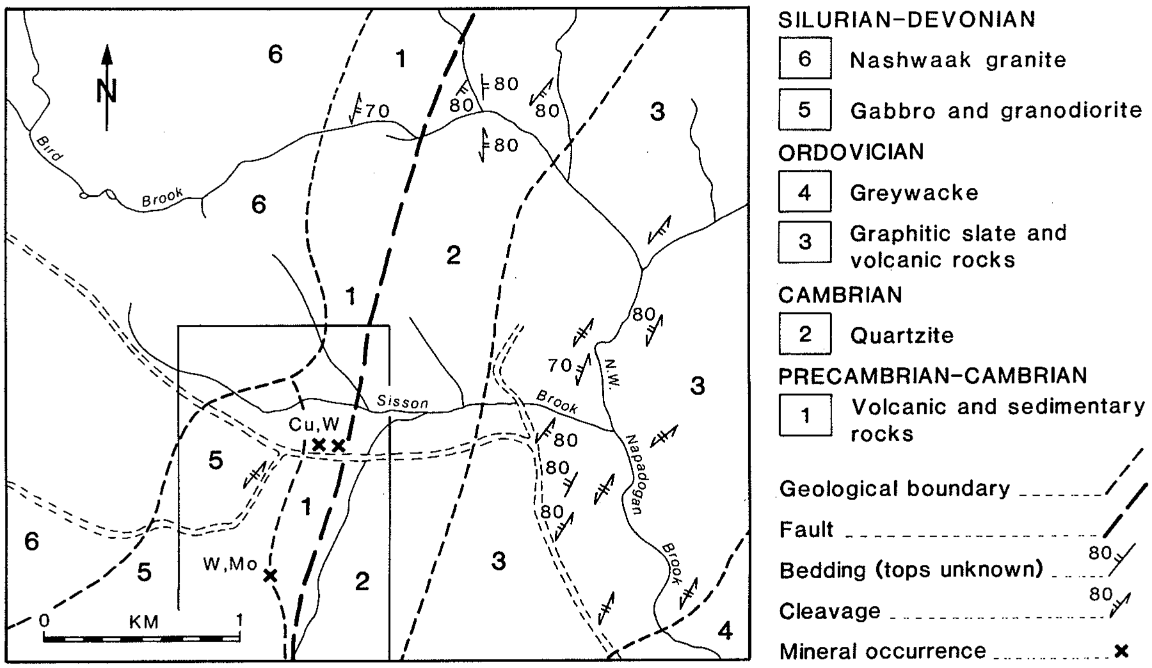


Figure 4. Geological map of the Sisson Brook area (modified after Lutes, 1981). The block outlines is shown in more detail on Figure 5.

associated with kaolin (white clay) and purple hematite alteration.

Greisen (type 5) mineralization is extensively developed in all but the Dungarvon Granite where it is confined to the northeast contact zone only. Type 5a veins contain wolframite (altering to scheelite), molybdenite and/or magnetite. The greisen contains cassiterite, although generally only 100-300 ppm Sn. Type 5bi veins have been noted to contain Cu-sulphides and molybdenite. Type 5bii veins contain coarse cassiterite and various sulphide minerals. Type 5c contains Cu, Pb, Zn, As, Sn, W, F, Be, and Mo in geochemically elevated but subeconomic concentrations. Type 5a mineralization is present at Peaked Mountain, Rocky Brook, Trout Brook and Tin Hill; type 5b mineralization is present at Trout Lake; and type 5c is present at the Trout Lake and Cleveland occurrences.

Types 6 and 7 sulphide showings are primarily Cu, Mo $\pm$ Sn-sulphides as disseminations and veinlets associated with grey-green (greisen ?) or brick red hematite alteration. Type 6a mineralization is found at the Little Dungarvon, Youngs Dam and Sisters Brook occurrences; type 6b at the Trout Lake occurrence;and type 7 at the Little Dungarvon occurrence.

Mineralized breccias (type 8) are of two types. Type 8a are hydrothermal in origin. They are pebble dykes with rounded, somewhat altered granite clasts in a quartz matrix. Some clasts are 100%

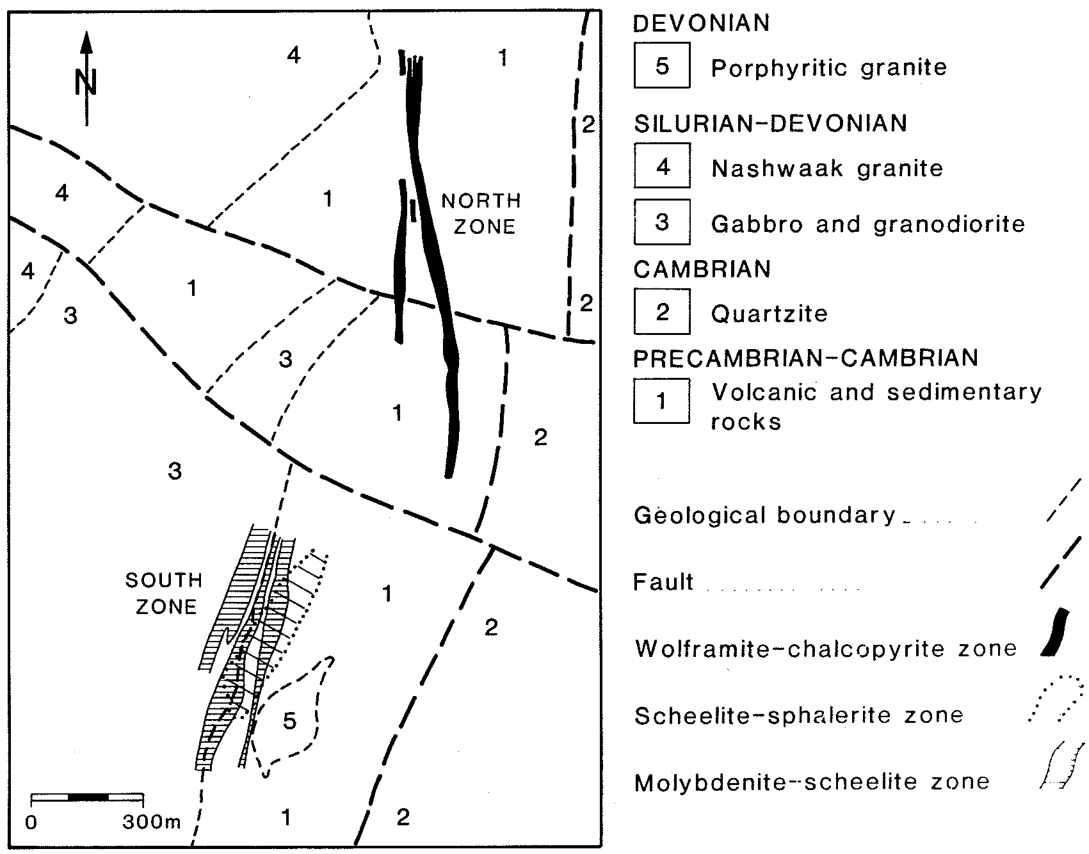


Figure 5. Geological map of the Sisson Brook tungsten-copper-molybdenum deposits (modified after Nast and Williams-Jones, 1984 and Enns, 1984).

pyrite. Fluorite is often found in the matrix material. Type 8b are in fault and/or tension zones which have opened up and been flooded with quartz, chalcedony and/or fluorite. Clasts are of metasedimentary rocks or granite and are angular and matrix-supported. Uranium may occur in some of these breccias. Types 8a and 8b are represented by the Carson Lake and Harris Lake occurrences respectively.

Of the exogranitic type 9 deposits (scheelite/molybdenite/molybdoscheelite/ quartz), stockwork veins (9b) are the most likely to produce economic deposits. Type 9a is typified by the Peaked Mountain deposit which consists of parallel veins of quartz cutting mafic volcanic rocks and intrusions. The veins have wide biotite-rich alteration selvedges. The area of best mineralization is approximately 600 by 200 metres. Type 9b is typified by the south zone at Sisson Brook (Fig. 4 and 5) where vein development is more of a stockwork with several vein orientations. This deposit is more extensive in that mineralization extends over an area of at least 2000 by 500 metres. This deposit also contains other metals such as Cu, Bi, and Zn.

Type 10a sheeted veins (exogranitic) contain W, Zn, Cu, Sn, Be, Mo and/or F in relatively closely spaced parallel veins with an average width of 10 cm at Burnthhill, 0.5 cm at McLean Brook South, (Fig. 6) 0.2 cm at Lightning Hill and 0.1 cm at Todd Mountain, all striking 120-140°. Type 10b are solitary veins, primarily narrow (1-2 cm) quartz-wolframite veins. Type 10c are wide (3-15 metres) silica-flooded zones which appear to be fault controlled. The north zones at Sisson Brook are of this type and contain wolframite (+ minor scheelite), Cu, Ag and Bi.

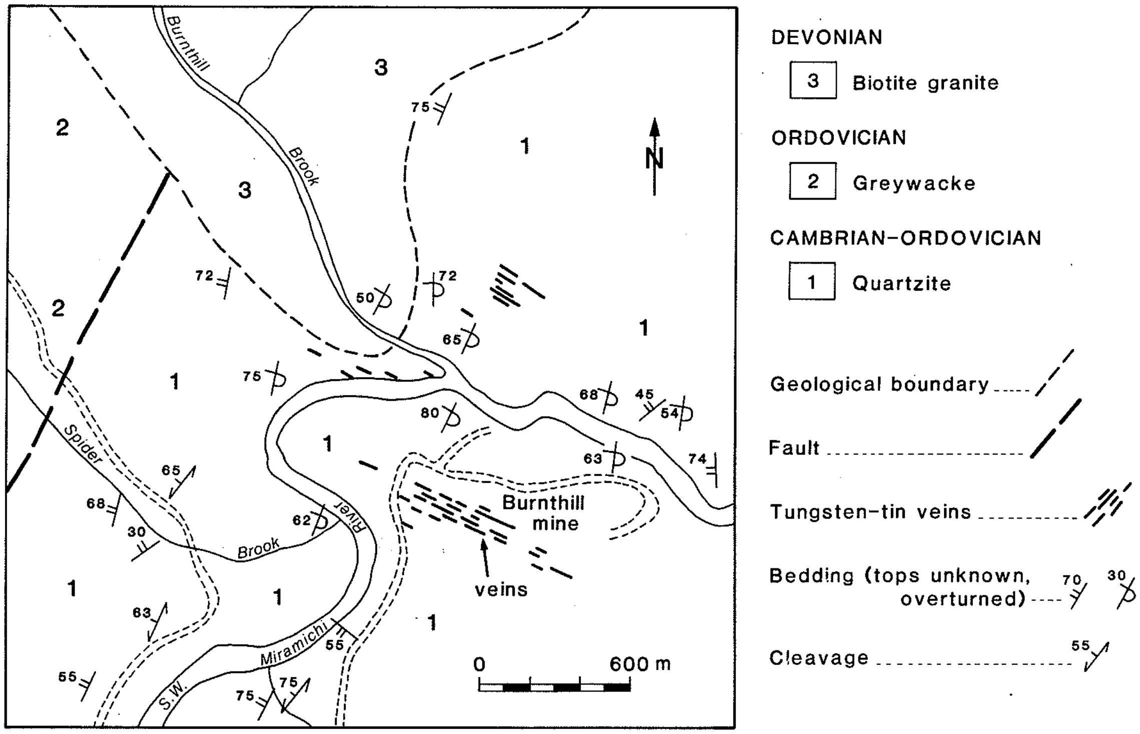


Figure 6. Geological map of the Burnthill Mine area (modified after Potter, 1969 and Crouse, 1978).

The type 11 multimetal veins appear to be the most distal from the mineralizing granites. They are characterized by zones of strong chloritic alteration with high Cu, Zn, Pb, Sn, W, F and Au values. Type 11 mineralization is present at the Lower Hayden Brook occurrence.

Type 12 veins are similar to type 1 veins but occur within the metasedimentary package. To date only fluorite and topaz has been found in these veins. One example of this vein type was found south of the Peaked Mountain deposit.

Type 13 occurrences are granitic dykes cutting hornfels which locally contain disseminated pyrite.

Skarn/hornfels deposits (type 14) are not well developed in the area due to low carbonate content of the rocks. Two recognizable deposits are type 14a calcareous siltstone horizons and type 14b mafic volcanics. The best example of 14a is a 60- metre wide calcareous siltstone near Lightning Hill which has been altered to quartz-epidote-feldspar ± diopside hornfels with a tin content of 120-450 ppm along a 2 km strike length. Skarn/hornfel zones in mafic volcanics contain epidote, Ca-garnet, actinolite and diopside. Tin contents range from 80-2000 ppm Sn. The tin is presumed to be in the calc-silicate minerals. Samples of 14a contain chalcopyrite and pyrrhotite while samples of 14b contain native Ag and Bi.

ROAD LOG: At the core shed of Canadian Pacific Forest Products Ltd. in Boiestown, there will be a display of drill core and samples from most of the mineral occurrences in the Burnthill area.

Following the core and sample display, field stops will be made to the Little Dungarvon area showings to view four types of tin mineralization as well as several phases of the Dungarvon Granite. From Boiestown, access to the area is by way of logging roads northwards from Holtville, a small village north of Boiestown. The Little Dungarvon showings are on Canadian Pacific's Freehold Grant and can be accessed only by passing through company gates. Field trip stops are illustrated on figure 7.

STOP 1A

Several feldspar veins (Vein 4 system) cut the coarser-grained phase of the Dungarvon Granite (2a). These veins vary from 30 cm to several metres in width. Vein material containing numerous vugs infilled with clear to faintly purple fluorite is exposed in the trenches. In the vicinity of the trench are boulders which also contain red-brown cassiterite in the vugs and as disseminations up to 1 cm in diameter. The best assay from these boulders was 28.8% Sn over 14 cm.

STOP 1B

Two tin veins (Vein 11 system) within 2 metres of each other cut the medium-grained phase of the Dungarvon Granite. The veins are quartz-feldspar veins which assay up to 10% Sn over 25 cm and in the trench grade 2.12% Sn over 2 metres. Drilling and trenching suggests that these veins grade into feldspar veins at depth and along strike to the southeast. The black cassiterite in these veins is generally fine-grained and shows some evidence of crushing. The granite here has weathered and exfoliated to the point that all constituent minerals are no longer bonded together.

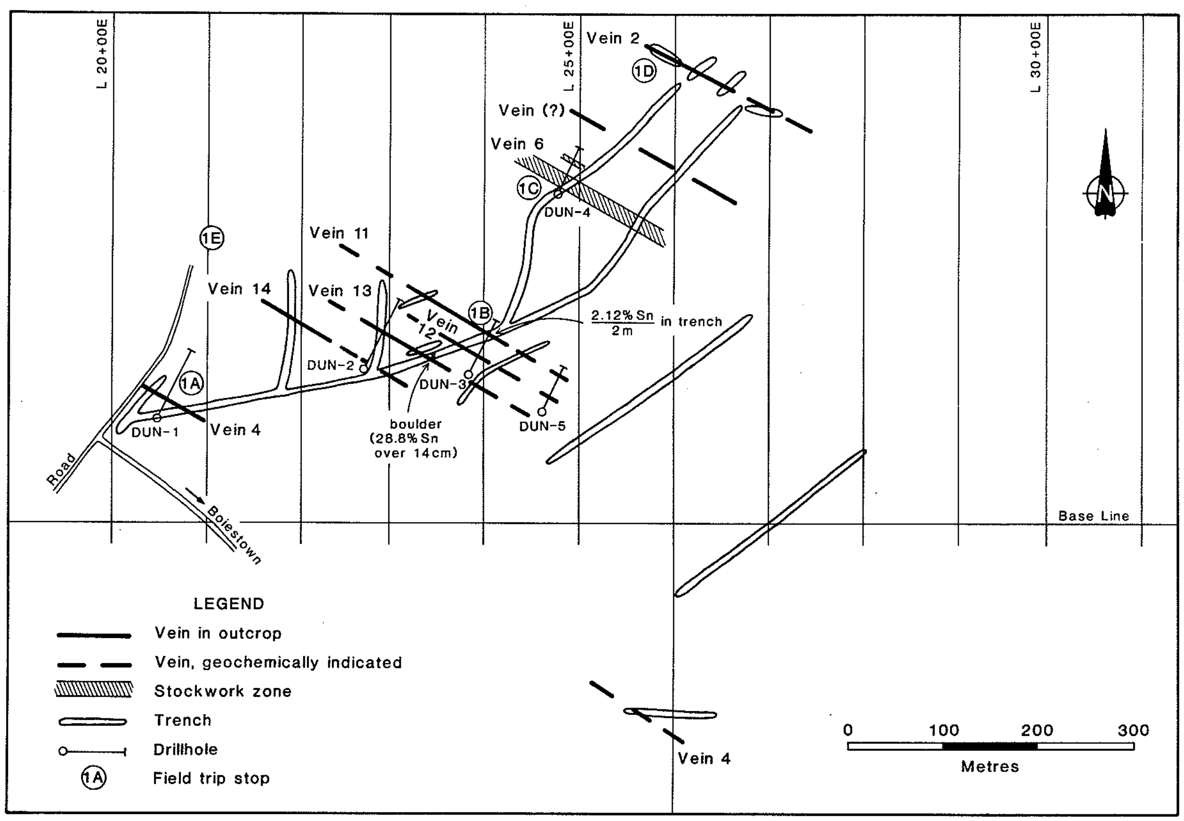


Figure 7. Tin-bearing veins and stockworks in the Dungarvon Main Showing area.

STOP 1C

The Vein 6 system is a 20-metre wide stockwork system dipping 45° to the southwest. Individual veinlets dip vertically or 60° to the northeast. The quartz ± feldspar ± cassiterite veinlets are 1-10 mm wide and occur at a density of 10-25/metre. A two-generational feldspar-cassiterite vein is also seen on surface but not in the drillhole. The zone grades 0.044-0.049% Sn over the 20 metres. The zone occurs within the intensely weathered granite and stands out as a low ridge of silicified and hematite-altered granite. The granite hosting the stockwork zone is the fine-grained phase of the Dungarvon Granite.

STOP 1D

The vein system is similar to that of Vein 6 with a poorer development of veinlets. This zone is only 2 metres wide and is vertical. Alteration here is siliceous only with no strong red hematite development.

STOP 1E

This outcrop is of one of the porphyry bodies which cut the main phases of the Dungarvon Granite. These porphyries occur as dykes and small plugs in this area. The porphyries primarly contain feldspar phenocrysts although quartz and biotite phenocrysts occur sporadically in minor amounts.

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GEOLOGY OF THE LAKE GEORGE MINE, SOUTHERN NEW BRUNSWICK[1]

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Introduction

The Lake George deposit is located near the village of Prince William, approximately 35 km southwest of Fredericton, New Brunswick (Fig. 1). This mine began production in 1970 and has since become a major producer of antimony in North America. In 1970, production began at the eastern margin of the north-dipping Hibbard vein which to date has yielded approximately one million metric tons of ore grading 3.0 to 3.5% Sb. In 1981, a second orebody was delineated downdip from and to the west of the existing mine workings on the same Hibbard vein. The estimated size of this orebody is approximately 0.8 million metric tons with average grade of 4.15% Sb. From drilling records, the shallow-dipping, east-trending Hibbard quartz vein system is now known to extend continuously to depths of at least 1 km and over a strike length of several kms.

The antimony ore in the Lake George mine occurs as lense-shaped tabular bodies that are irregularly distributed in north- to northeast-raking zones within the east-trending Hibbard quartz vein system. The remaining parts of this vein system, although mineralized, are either diluted in grade or too narrow in width to be economic. Other quartz veins in the mine area contain only minor amounts of antimony.

During the 1981 exploratory drilling program, an extensive zone of scheelite- and molybdenite-bearing calc-silicate and quartz veinlet-stockworks was outlined both in the hanging wall and the footwall of the Hibbard vein (Seal II et al., 1987). This W-Mo stockwork, spatially related to the periphery of a buried monzogranite cupola, occupies steeply-dipping fractures.

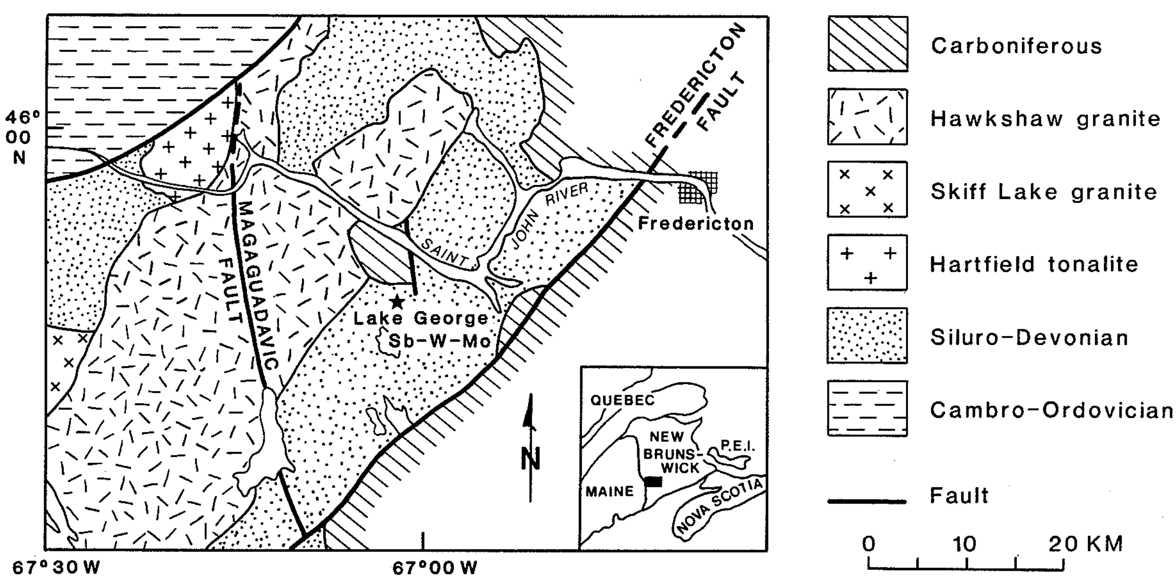

Figure 1. Geological setting of the Lake George Mine (modified after Seal II et al., 1988).

[1]This paper is based largely on publications by Scratch et al. (1984), Seal II et al. (1987, 1988), Morrissy and Ruitenberg (1980) and Morrissy et al. (1985). It was published previously in a Geological Association of Canada guidebook by Procyshyn et al. (1989) and is reprinted here with minor revisions.

Underground drifting associated with the pre-production development of the second antimony ore body exposed a third style of mineralization in the footwall of the Hibbard vein. This mineralization contains sphalerite-rich stibnite-calcite-quartz veinlets that are associated with spectacular bleach patterns in calcareous host rock. These veinlets also contain minor galena, arsenopyrite, bismuthinite, native bismuth and native gold as well as minor scheelite (Seal II et al., 1987). Arsenopyrite- and locally pyrite-enriched areas, which occur as dark alteration margins to pervasive bleach alteration fronts, are reported to carry gold values.

The main antimony ore zone, at least in part to the southeast corner of the older workings, is enveloped by a uranium-bearing zone that contains thucholite, uraninite, coffinite and fluorite. It is generally felt that these radioactive zones are associated with late phases of mineralization that host the most notable concentrations of native antimony in the form of irregular masses that also contain stibnite and pyrrhotite, and veinlets which locally transect the irregular masses of native antimony, stibnite and pyrrhotite. These concentrations of native antimony and U-bearing minerals are generally restricted to the hanging wall of the main antimony ore zones. Native antimony also generally predominates in the western part of the main quartz-stibnite Hibbard vein system (Morrissy and Ruitenberg, 1980).

The Lake George deposit thus represents metal concentrations within a complex polymetallic hydrothermal centre (Seal II et al., 1988). The age of this deposit is not certain, but preliminary K-Ar age determinations on micas (Seal II et al., 1988) suggest that the mineralizing events were closely associated in time and occurred during the Late Silurian (411-417 Ma) Period. This age is problematic since these deposits in large part clearly post-date the intrusive phases that structurally and compositionally appear to be related to the Pokiok Batholith which is most probably Early Devonian in age (McCutcheon et al., 1981).

Geological Setting

The Lake George deposits are hosted within deformed Silurian turbiditic metasediments of the Fredericton Cover zone (Fyffe and Fricker, 1987). The massive to well-bedded metasedimentary units contain greywacke-sandstone, siltstone and black, non graphitic slates. All units contain varying amounts of carbonate but massive limestone horizons have not been observed. Structurally, the rock units, now tightly folded and metamorphosed to lower greenschist facies, are penetratively cut by a northeast-trending axial planar cleavage that is related to the Acadian orogeny (Ruitenberg and McCutcheon, 1982). Subsequent to the orogeny, the area was intruded by the polyphase Pokiok batholith of probable early Devonian(?) age. The Hawkshaw monzogranite is the youngest phase of this batholith and outcrops approximately 3 km northwest of the mine site. The margins of this pluton are coarse-grained and porphyritic. The megacrystic monzogranite grades inward to a younger medium grained, equigranular muscovite-biotite granite. The monzogranite contains quartz, zoned plagioclase, perthitic alkali feldspar (microcline and orthoclase), biotite, sporadic muscovite and rare hornblende. Chemically, the monzogranite straddles the plagioclase-biotite tie-line on the ACF diagram and its low ferric iron ratio is indicative of the ilmenite series of granite that is transitional between I- and S-type granites (Seal II et al., 1988). The monzogranite bodies and the Silurian metasediments are unconformably overlain by essentially undeformed fluviatile sediments and minor volcanics of the Carboniferous cover sequence exposed in a half graben north of the mine site.

There is present at surface, in the vicinity of the Lake George deposits, two different dyke sets (Fig. 2). The earliest of these are east-west-trending and steeply south-dipping lamprophyric dykes, 0.5 - 4m thick. These alkali-rich (K + Na) shoshonitic lamprophyre (Seal II et al., 1988) dykes are invariably altered in the vicinity of the deposits. The second dyke set is a north- to northwest-trending and steeply west-dipping quartz-feldspar porphyry dyke system that has also been traced across the entire mine property. These dykes are generally 3 to 5 m thick. At their most northerly extremities, the main dykes break up into several parallel dykes that have lens-like forms elongate along the main trend. The phenocrysts of quartz are commonly resorbed, and those of plagioclase are

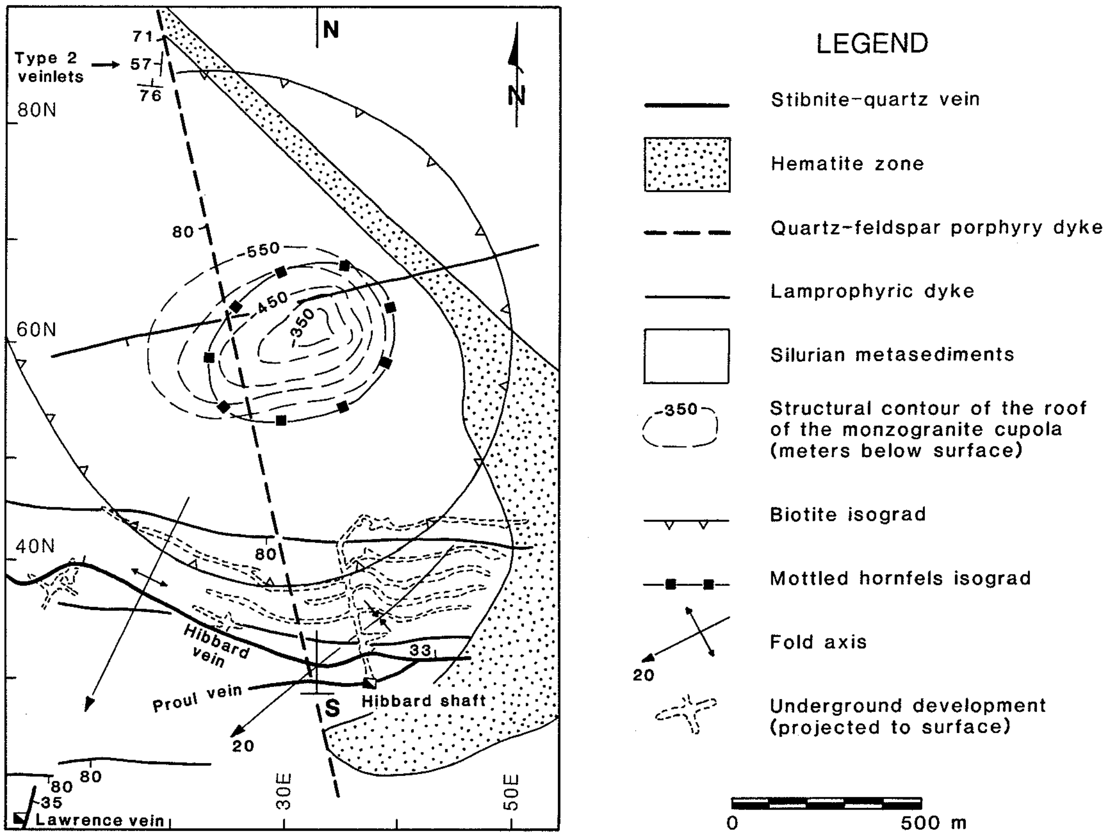

Figure 2. Schematic surface geology in the vicinity of Lake George Mine (modified after geological mine maps, Seal II et al., 1988 and Scratch et al., 1984). Location of section on Figure 3 is shown as N-S.

strongly zoned. These rocks texturally grade to seriate granularity of monzogranite composition. The lamprophyre dykes are cut and offset by the porphyry dykes. Both the lamprophyre and porphyry dyke systems are crosscut and altered by the Hibbard Sb-quartz veins; however, the scheelite-molybdenite quartz stockwork in part predates the intrusion of the porphyry dykes and in part postdates their emplacement.

Diamond drilling and geophysical data have outlined a completely buried monzogranite cupola 500 m north of the Hibbard shaft at a depth of 435 m. The monzogranite is fine- to medium-grained and texturally seriate. It contains partially resorbed quartz, zoned plagioclase, and biotite in a finer grained matrix of K-feldspar and trace amounts of hornblende, apatite and sphene. These monzogranites are compositionally similar to the porphyry dyke and to the Hawkshaw pluton. Both the cupola and the felsic dykes greatly resemble, both in texture and composition, certain porphyrites in the Mt. Pleasant complex (Morrissy and Ruitenberg, 1980).

The cupola is surrounded by a well-defined thermal aureole that clearly is superimposed on the greenschist facies of regional metamorphism. Isograds in pelitic rocks have been delimited (Scratch et al., 1984) on the basis of the first appearance of (i) biotite, (ii) a mottled or spotted hornfelsic texture characterized by clots of muscovite, quartz, chlorite and biotite presumably retrograded from cordierite xenoblasts and andalusite porphyroblasts (Seal II et al., 1988) and (iii) a narrow zone adjacent to the intrusive contact containing the assemblage corundum + K-feldspar. The assemblage corundum + K-feldspar, in the stability field of andalusite, would place an upper pressure limit on contact metamorphism of ca. 175 MPa (Seal II et al., 1987).

In rocks of marly composition, three additional isograds have been mapped on the basis of the first appearance of (i) tremolite, (ii) the assemblage tremolite + K-feldspar and (iii) diopside. Furthermore, much of the calc-silicate hornfels contains the assemblage clinozoisite-plagioclase-calcite which can be used to buffer the $H_2O:CO_2$ ratio in the rocks (Greenwood 1975). Using these observations and microprobe mineral composition data, Seal II (1984) from his thermochemical calculations, concluded that the isograds in the calc-silicate rocks were produced at lower temperatures than those recorded within the adjacent pelitic rocks. The calc-silicate assemblages record, therefore, conditions that had occurred not for peak metamorphism but for retrograde metasomatism during which the cordierite and andalusite were altered to their present assemblages. These calc-silicate isograds, while being essentially parallel to the isograds in pelitic rocks on the north side of the cupola, are strongly discordant to the isograds in the pelitic rocks on the south side of the cupola.

The cupola is cut and altered by the stibnite-quartz veins and by the quartz-scheelite-molybdenite veinlets.

Structural Geology

The regional structures in the area are characterized by tight folds and the presence of a penetrative northeast-trending axial-planar cleavage that developed during the Acadian orogeny. At the mine site, an antiformal/synformal fold pair trends N 20° E to N 50° E, plunges 20° SW and has an axial-planar cleavage that dips 80° NW (Morrissy and Ruitenberg, 1980). A smaller second-order anticlinal fold immediately east of the Hibbard-Prout shaft is essentially a parasitic deflection developed on the limb of a regional syncline. All intrusive activity and mineralization in the area are now thought to postdate the folding event (Seal II et al., 1988) although earlier reports (Scratch et al., 1984) have suggested that the contact isograds and the Hibbard-Prout vein system may have also been folded about the same regional axes of deformation. The northwest deflection of the east-trending Hibbard vein on its westward extension, is most probably due to refraction of the vein fracture in response to the presence of more competent beds in the vicinity of the regional synform west of the Hibbard-Prout shaft.

Types of Mineralization

The mineralization at Lake George consists of 5 distinct associations: (i) a scheelite-molybdenite veinlet stockwork, (ii) a quartz-carbonate, locally sphalerite-rich, veinlet system containing native bismuth, bismuthinite and native gold, (iii) the main quartz-stibnite vein system, (iv) a late quartz-carbonate-stibnite-pyrrhotite-rich pervasive veinlet system carrying native antimony and locally fluorite with uranium-bearing minerals and (v) an extensive hematite-cemented breccia body that partially envelopes the eastern limits of the main Hibbard quartz-stibnite vein system. Only the scheelite- molybdenite system and the Hibbard quartz-stibnite system have been well studied and their areal extent outlined (Seal II et al., 1988, 1987 and Scratch et al., 1984 repectively)(Fig. 3). To date, only the quartz-stibnite vein system has been mined.

The W-Mo Mineralization

The W-Mo deposits comprise three different scheelite- and/or molybdenite-bearing veinlet types which can be distinguished on the basis of mineralogy, associated alteration selvages and relative age. The earliest veinlets (Type 1) which contain scheelite and molybdenite, are composed dominantly of calc-silicate minerals and lesser amounts of quartz. These veinlets clearly postdate the thermal metamorphic assemblages in pelitic hornfels, but as yet have not been observed to occur within the monzogranite cupola. Type 1 veinlets are cut by lampropyre and porphyry dykes. Type 2 veinlets containing scheelite and molybdenite are quartz dominant and have developed subsequent to emplacement of porphyry dykes. Type 3 veinlets containing molybdenite only in association with prehnite and quartz have developed subsequent to Type 2 veinlets but definitely prior to the stibnite-quartz veins.

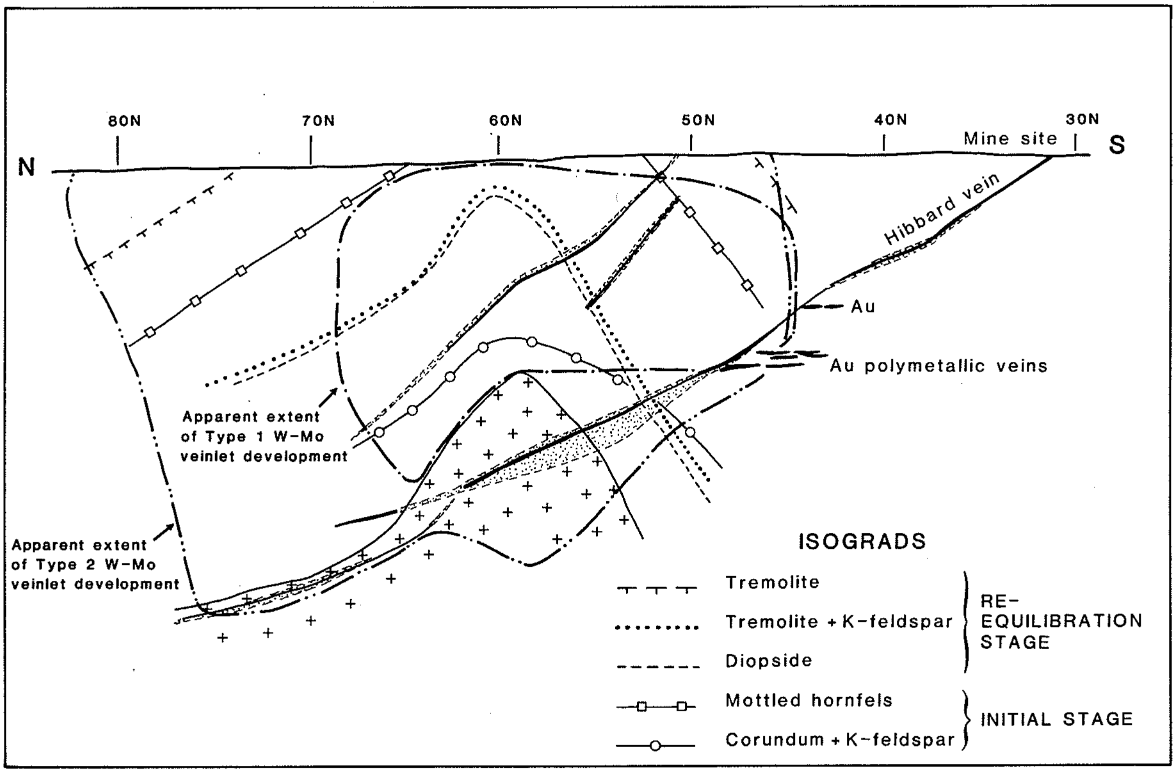

Figure 3. Composite north south section showing the various Lake George deposits (modified after Morrissy et al., 1985; Seal II et al., 1988).

Type 1 veinlets, ranging from 5 to 50 mm in width, contain granditic garnet, wollastonite, clinopyroxene and calcic amphibole hosted by minor amounts of intergranular quartz and calcite which tend to concentrate toward the cores of the veinlets. Wollastonite and garnet are commonly restricted to the cores of these veinlets, and almost always occur with lesser amounts of clinopyroxene and calcite. Clinopyroxene more typically is most abundant at the vein boundaries, but in the absence of wollastonite or garnet, can occur as isolated grains in the calcite-quartz cores. Scheelite, molybdenite and pyrrhotite are also locally present in the cores of these veinlets. Strongly pleochroic acicular grains of calcic amphibole locally replace the clinopyroxene. Clinozoisite, sphene and apatite also occur in minor amounts in the veinlets. The quartz dominated Type 1 veinlets contain lesser amounts of granditic garnet, clinopyroxene, calcic amphibole, plagioclase clinozoiste, calcite and sphene. Two varieties of calcic amphibole are present: (i) a strongly pleochroic, acicular variety replacing clinopyroxene, and (ii) a pale green weakly pleochroic bladed variety that appears to be primary in origin. The quartz-rich cores commonly contain scheelite, molybdenite, pyrite, pyrrhotite and chalcopyrite.

Type 1 veinlets are ubiquitiously enveloped by thin bleached metasomatic alteration margins that are most apparent in the darker hornfelsic hosts. In the alteration margins, the ferromagnesian minerals produced during the hornfelsing process, have been transformed to tremolite, plagioclase, sphene and, to a lesser extent, clinopyroxene and clinozoisite.

Type 2 W-Mo veinlets, ranging from a few mm to 15 cm in width, are dominated by milky quartz that contains subordinate amounts of alkali feldspar, calcite, muscovite, biotite, calcic amphibole and sphene and locally host scheelite, molybdenite, and pyrite. The alkali feldspar now contains twinned albitic plagioclase exsolved from a host that is dominantly orthoclase but which contains small zones of twinned microcline. Large (up to 1.5 cm across) subhedral to euhedral grains of scheelite and/or large euhedral booklets of molybdenite are commonly localized with sparse amounts of pyrite at the margins of these veinlets. Molybdenite also occurs locally throughout the veinlets. Where these veinlets cut the monzogranite of the cupola or pelitic hornfels, their margins are lined by subhedral booklets of muscovite and biotite whereas those veinlets cross-cutting calc-silicate hornfels more commonly contain subhedral to euhedral grains of calcic amphibole and sphene at their margins. Muscovite also coexists locally with K-feldspar and quartz in the veinlets cutting the cupola.

The alteration envelopes associated with Type 2 W-Mo veinlets are variable and depend on the nature of the host rock. In monzogranite, alteration involved sericitization and carbonatization of the feldspar and chloritization of biotite for distances up to 2 cm into the wall rock. In these instances molybdenite, occurring as fine grained booklets, is intimately intergrown with calcite mosaics in the wall rock. In pelitic rocks the only observable alteration is the chloritization of the biotite and local formation of sericite or plagioclase or calcite for distances less than 2mm into the wall rock. In calc- silicate hornfels, the only observable change is the complete removal of clinopyroxene and clinozoisite from the wall rock.

Along the northern side of the cupola and approximately 800 m away from the intrusive contact, the calc-silicate hornfels contains a pervasive form of W-mineralization in the form of anhedral poikilitic grains of scheelite containing fine grained quartz, calcite, tremolite, plagioclase, clinozisite, sphene and ilmenite. This form of mineralization often occurs adjacent to type 2 veinlets and is believed part of this same event.

Type 3 veinlets, usually less than 2 mm in width, contain prehnite, quartz and molybdenite, with minor amphibole, calcite and chlorite. Alteration envelopes can be seen only where veinlets crosscut pelitic hornfels. Adjacent to the veinlets, the sodic plagioclase and biotite in the hornfels across microscopic selvages, are transformed to intermediate to calcic plagioclase and calcic amphibole respectively. Type 3 molybdenite veinlets are only sparsely and erratically developed.

Distribution of Types 1 and 2 W-Mo Veinlets

Type 1 mineralization is broadly concentric over the apex of the monzogranite cupola. In general, garnet and clinopyroxene tend to occur closer to the monzogranite contact than do their hydrous counterparts, clinozoisite and calcic amphibole. Scheelite and molybdenite are also more abundant in the veinlets proximal to the cupola. The veinlets are both concordant and discordant with respect to bedding. Their orientation is unrelated to regional structures and probably is related to hydraulic fracturing associated with the evolution of the vapor phases from the monzogranites (Seal II et al., 1988).

Type 2 mineralization is volumetrically more significant than that of Type 1 and thus considered to have a greater economic potential. This form of mineralization is observed at least 750 m north, 200 m south, 250 m east and 400 m west of the apex of the cupola with the stronger mineralization tending to occur north of the cupola apex. To date only local pockets over a few 100 m have been delineated where grades exceed 0.4% WO_3.

Polymetallic-Gold Mineralization

As indicated earlier, this mineralization type, often associated with significant gold content, has only recently been recognized in footwall drifts that were cut during the development work on the second antimony ore body. The veins are dominated by quartz, calcite and stibnite that occur in association with sphalerite, pyrrhotite, native bismuth, bismuthinite, galena, bornite, chalcopyrite, pyrite, arsenopyrite, native gold and tetradymite. Scheelite is also usually present in the veins as a minor phase. Sphalerite is locally a dominant phase. In the vicinity of the veins, the calcareous metasedimentary rock units often display a characteristic pervasive and locally extensive zonal greyish-green alteration pattern, in which arsenopyrite and lesser amounts of pyrite appear concentrated in distinct fronts. Both these minerals carry gold values.

When hosted by pelitic hornfels, these shallow easterly-dipping and northwest-trending quartz-calcite veins display zoned alteration selvages of considerable thickness that often exceed the vein widths. These alteration envelopes are mineralogically dominated by sericite (up to 85% of the rock mass) but also contain quartz and minor chlorite. The colour banding in these envelopes is parallel to the vein boundary and only reflects differences in the abundance of disseminated arsenopyrite and pyrite.

The quartz-carbonate veins are best exposed in the footwall of the second Sb-ore body within the Hibbard quartz-stibnite vein, where they occur in swarms that contain up to ten veins across zones 6 to 12 m wide. Here, the polymetallic-Au bearing veins postdate the W-Mo mineralization, but in turn are crosscut and locally offset by the main quartz-stibnite vein.

Stibnite-Quartz Mineralization

The antimony ore in the Lake George mine forms lens-shaped bodies which are irregularly distributed as northeast- and locally north-raking zones within an east-trending quartz-vein system that has been traced over 1 km in depth and for over 2 km along strike length. This vein system is known as the Hibbard vein and dips 30° to 35° to the north. The Prout vein, a minor past producer, is a west-southwest-trending splay of the main Hibbard vein system which merges eastward with the main vein system and shortens with depth so that it no longer is present below the fifth level. Additional east-trending quartz-stibnite veins are known in the mine area which are subparallel to the Hibbard vein system but these to date have proven to be subeconomic. The Hibbard vein within the ore zones averages 1.0 to 1.5 m in thickness and contained approximately 1 million short tons of ore grading 3.0-3.5 wt% Sb. Approximately 80% of the ore mined from this orebody came from one ore shoot in the central and eastern mine workings. More recently a second orebody was delineated donwn dip from and to the west of the existing mine workings. This orebody is estimated to contain 0.8 million tons of ore grading 4.15 wt.% Sb (Ellis, 1983). The Prout vein averaged 0.3 m in thickness and contained approximately 42,000 tons of ore grading 2.42 wt.% Sb (Scratch et al., 1984).

The second quartz-vein system on the property strikes north-south and dips 35° to the east. This vein system however, is subeconomic. Some production has been derived from one such vein called the Lawrence vein but the amount, although uncertain, was definely much less that that extracted from the Prout vein.

Two additional shafts on the property, the Adams Shaft and the Hibbard-6 Shaft, were developed on the western extent of the Hibbard vein.

Structurally, the two main ore bodies are localized within the Hibbard vein system where they transect the axial regions of major northeast-trending regional folds. These folds plunge approximately 20° to the southwest. In the axial regions, the vein system developed minor flexures in its trend and increased dilation due to more pronounced refraction of the vein trend in response to increased apparent thickness of the more competent beds in these localities. The now depleted Number 1 orebody was localized on a synformal axial region whereas the number 2 orebody located west of and downward of the first orebody, is localized in the vicinity of an antiformal axial region. Part of the No. 2 orebody is contained within the monzogranite cupola.

Locally there is evidence for right lateral movement along the antimony-bearing quartz veins that occupy a fracture zone. Repeated movement along this fracture zone is especially evident where the Hibbard vein traverses the axial region of the regional folds. Thus, intense brecciation and repetitive healing by stibnite mineralization and darker gray quartz characterize the trough-shaped vein in the axial portion of regional folds. In these areas, the stibnite deposited early in the history of the vein is characterized by the development of twin planes, which produced glide surfaces closer to the fold axes. At the hinges of the folds, stibnite has been recrystallized to a fine grained, equigranular aggregate of strain-free grains. On the flanks of the open folds, stibnite within the vein is undeformed and occurs as euhedral crystals up to 2 cm long, and show no development of intracrystalline strain features. The older workings extend, with depth, into a siliceous hornfels zone that contains abundant quartz stringers with stibnite and chalcopyrite blebs (Morrissy and Ruitenberg, 1980).

Scratch et al. (1984) attributed the sinuous configuration of the Hibbard vein to post-vein folding. The evidence for this interpretation is weak. The emplacement of the monzogranite cupola and associated metamorphic aureole and all styles of mineralization in the area, clearly postdate the development of the penetrative axial plane cleavage related to regional folding. Unlike the earlier formed W-Mo veinlets, however, whose orientation and spatial association can be related to the emplacement of the cupola, the development and orientation of the stibnite-quartz vein appears to have been controlled by later stresses of a regional nature (Seal II et al., 1988).

The main ore-forming minerals at the Lake George mine are stibnite and native antimony. In the eastern part of the mine stibnite and quartz strongly dominate the mineral composition of the No. 1 ore body, whereas in the western workings near the Adams shaft, native antimony with calcite are also important components in the ore veins. Lesser amounts of pyrite, pyrrhotite, arsenopyrite and tetrahedrite containing exsolution blebs of chalcopyrite are also present in the ore. Trace amounts of bournonite ($2PbS\ Cu_2S\ Sb_2S_3$), chalcostibite ($Cu_2S\ Sb_2S_3$), plagionite ($5PbS\ 4Sb_2S_3$) and fuloppite ($3PbS\ 4Sb_2S_3$) are present within lead-rich portions of the orebody (Abbott and Watson, 1975). The main uranium minerals identified in the radioactive zone surrounding portions of the orebody, which are usually rich in native antimony, include thucholite, uraninite and coffinite with trace amounts of soddyite and uranophane (Scott, 1979). Muscovite and albite have also been identified in the quartz veins.

In the No. 1 orebody, stibnite and pyrite are dominant in the eastern workings, whereas native antimony and pyrrhotite are dominant in the western portion of the orebody (Scratch et al., 1984). The No.2 orebody is, however, essentially unzoned. Minor amounts of native antimony and pyrrhotite occur throughout the vein in association with stibnite, pyrite and lesser amounts of arsenopyrite. These associations can be observed even at polished section scale. The fact that native antimony has been observed in direct contact with pyrite possess interesting phase relationships (Ellis, 1983) that cannot be immediately resolved.

The stibnite-bearing quartz veins are everywhere enveloped by mineralogically distinctive alteration assemblages called argillic facies (Scratch et al., 1984) (or phyllic facies by Seal II et al., 1988) and siliceous facies. The yellow-green-coloured argillic facies zone is characterized by an ultrafine grained assemblage of quartz, sericite, illite, kaolinite, and dickite that host finely disseminated euhedral to subhedral grains of arsenopyrite, pyrite, pyrrhotite, chalcopyrite, and minor magnetite. Rare tetrahedrite, tenantite and native gold have also been identified in these alteration zones. Stibnite, although occasionally seen to rim arsenopyrite in the argillic alteration areas, is a later form of mineralization that clearly crosscuts these alteration zones. The argillic zone, exhibiting a knife sharp contact with pelitic hornfels, alters both the cordierite and biotite in the contact metamorphic aureole. Chemically, the wall rocks affected by the argillic alteration appear to have been depleted in Fe, Mg, Ca and Na and enriched in Si and generally K (Scratch et al., 1984).

The siliceous facies zone forms a narrow grey coloured inner band approximately 1 to 5 cm thick immediately adjacent to the main stibnite-bearing quartz veins and is overprinted on the previously-formed outer argillic facies zones. These silica-rich bands consist primarily of fine grained quartz that has completely replaced the matrix but not the primary rounded detrital quartz grains up to 0.5 mm in diameter that locally form up to 50% of these bands. Minor amounts of arsenopyrite, pyrrhotite, pyrite and stibnite can be seen in these siliceous bands. The siliceous alteration bands are coextensive and contemporaneous with the stibnite-quartz vein. No crosscutting relationships has ever been observed between the alteration bands and the veins.

A less well documented alteration type which lies outside the argillic alteration envelope of the Hibbard vein, is apparently characterized by the development of hydrothermal biotite that yields a reddish-brown colouration in the metapelitic rocks (Seal II et al., 1988). Ellis (1983) concluded that this biotite zone is intermediate in chemical composition between unaltered metapelite and the phyllic zone with respect to the Fe, Mg, Na and Si contents.

Hematite Breccia Zone

The hematite breccia zone occupies a broad area (approximately 1 km^2) that envelopes the eastern limits of the Hibbard vein. The hematite breccia zone is an intensely shattered metapelitic rock that is cemented by hematite, quartz, magnetite and minor amounts of pyrite and chalcopyrite. The eastern limits of the stibnite-bearing quartz veins are cut by and in turn transect the hematite-rich veinlets, as it anastomoses, thins and terminates with rock containing the hematite breccia. Drilling records indicate that the stibnite-quartz veins do not persist within the zone of hematitic breccia. The hematite-associated mineralization clearly overprints the cordierite and biotite contact metamorphic aureole as well as the argillic alteration envelope that surrounds the stibnite-quartz veins. The hematite breccia mineralization appears to be contemporaneous with, but spatially and chemically distinct from, the stibnite-quartz veins and its attendant siliceous alteration envelope (Scratch et al., 1984). To date this unusual zone has received little study.

Deposit Genesis

Detailed studies of the mineralogy, alteration and fluid inclusions of the Lake George deposit (Scratch, 1981; Seal II, 1984) have resulted in different interpretations. There is a general concensus among the authors that the scheelite mineralization was related to the emplacement of the monzogranite intrusion. Fluid inclusion data from the scheelite-bearing assemblage suggest these minerals were deposited from brines of moderate but variable salinity (ranging from 4 to ca. 20 equivalent weight % NaCl) at temperatures between 550° and 175°C (Seal II, 1984). The W-Mo mineralization is spatially and probably genetically related to the emplacement of the monzogranite cupola.

Several different models have been proposed to explain the genesis of the antimony-bearing quartz veins. Scratch et al. (1984) concluded that the veins were formed between 325° to 350°C from a boiling solution at a depth of approximately 1.5 km below the prevailing water table. They suggested that the monzogranite intrusion produced the fracture system and caused convection of saline fluids that leached Sb and Au from previously enriched shales. The hematitic breccia zone was considered to be the recharge area for the hydrothermal cell. Their second model suggests assimilation of the country rocks by the monzogranite which also generated the hydrothermal fluids during crystallization. In the latter case, the hematitic alteration zone is considered to represent the encroachment of groundwater on the juvenile hydrothermal system. The highly anomalous antimony content of the enclosing rocks of the area support previous antimony enrichment of these rocks in the area. However, the antimony bearing quartz veins clearly cut across the monzogranite intrusion and spatially seem unrelated to its presence.

Seal II et al. (1988) re-examined the fluid inclusions present in quartz and consider data obtained from the secondary inclusions in these quartz veins to be more representative of conditions present during deposition of antimony than do data from the primary inclusions. Furthermore, improvements in measuring technique and instrumentation have resulted in the recognition of clathrates and CO_2 vapour in the fluid inclusions. They attribute the variance in phase proportions in the fluid inclusions to immiscible separation of CO_2-rich fluids from saline hydrous fluids, and not directly to boiling as suggested by Scratch et al. (1984). Assuming that mineralization occurred at a confining pressure of 130 MPa as derived from calculations based on metamorphic assemblages present in the contact aureole of the cupola, and applying the necessary pressure corrections to the fluid inclusion data, Seal II et al. (1988) inferred that the quartz deposition occurred from 480°C to ca. 420°C and that stibnite was deposited at ca. 420°C to possibly as low as 300°C. The brine salinities for these fluids ranged from 1.5 to 18.0 equivalent weight % NaCl.

TOUR:

It is not possible to provide an underground tour of the Lake George Mine at the time of the excursion. Representative drill core sections will show the following styles of mineralization: (i) quartz-carbonate stringers containing scheelite and molybdenite that occur within the contact aureole of the monzogranite (ii) auriferous aresenopyrite rich alteration assemblages associated with quartz-carbonate veins, (iii) stibnite-bearing quartz veins surrounded by micaceous and argillic alteration envelopes.

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THE MOUNT PLEASANT CALDERA: GEOLOGICAL SETTING OF ASSOCIATED TUNGSTEN-MOLYBDENUM AND TIN DEPOSITS[1]

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Introduction

The Mount Pleasant caldera is a north-south trending elliptical feature with minimum dimensions of 13 by 34 km as outlined by regional gravity and magnetic maps. The northern half is concealed by overlying Middle Mississippian and Pennsylvanian strata. The caldera is bounded to the east and west by polydeformed Ordovician to Lower Devonian turbiditic metasedimentary rocks of the Digdeguash, Flume Ridge and Waweig formations. Late Silurian to Devonian granitic rocks of the Saint George batholith form part of the boundary along the southern margin of the caldera. Rocks within the caldera comprise part of the Upper Devonian (previously considered to be Mississippian) Piskahegan Group that is divisible into exocaldera, intracaldera and late caldera-fill sequences (McCutcheon, 1985, 1990 and Fig. 1). Each is described briefly below.

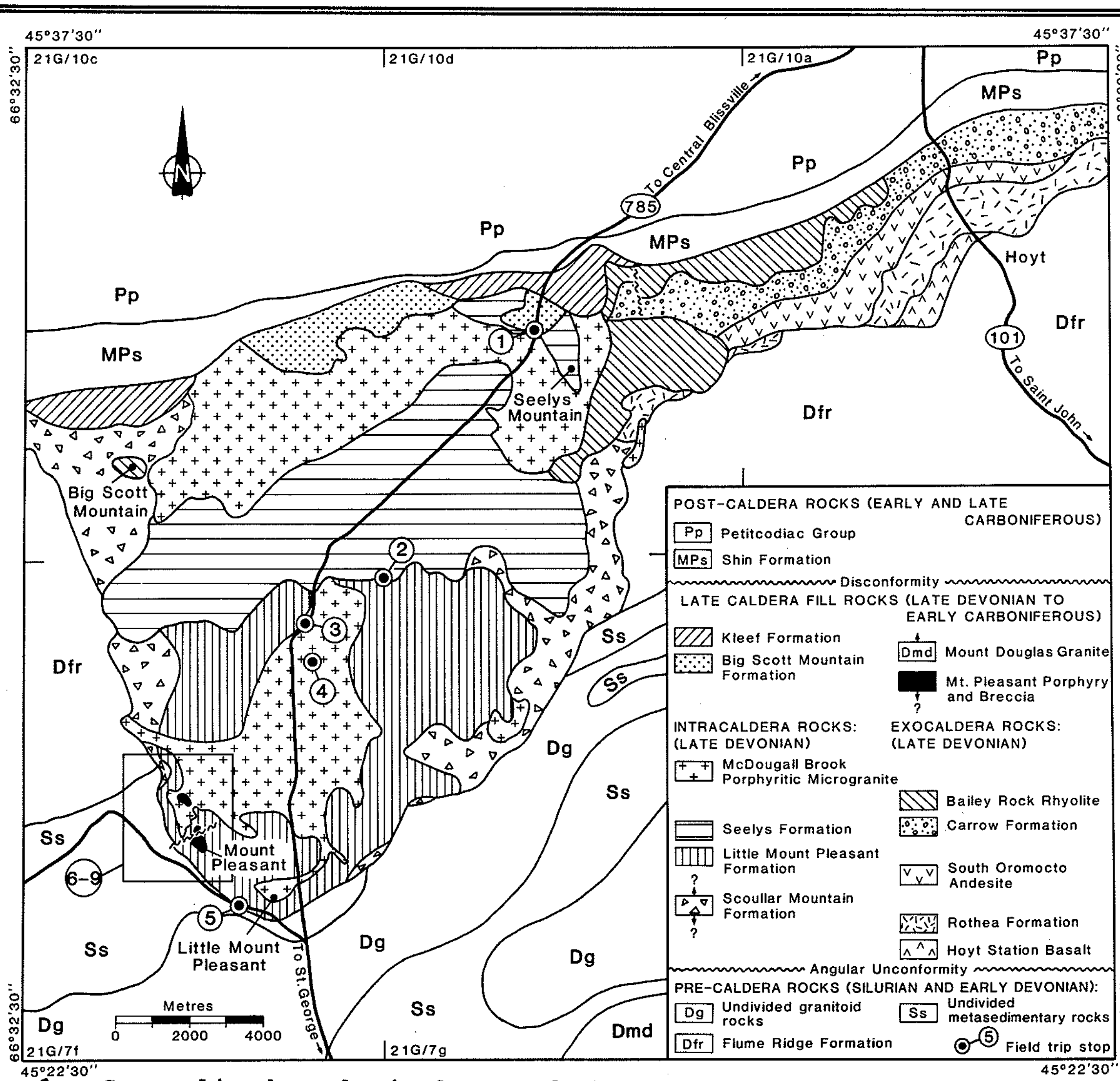

Figure 1. Generalized geological map of the Mount Pleasant caldera (modified from McGregor and McCutcheon, 1988).

[1]This paper was published previously in a Geological Association of Canada guidebook by Procyshyn et al. (1989) and is reprinted here with minor revisions.

The Exocaldera Sequence

The exocaldera sequence, in ascending stratigraphic order, consists of the Hoyt Station Basalt, Rothea Formation, South Oromocto Andesite, Carrow Formation and Bailey Rock Rhyolite. The first and last units have the least areal extent.

The Hoyt Station Basalt comprises at least two flow units. Minor pebble- to cobble-conglomerate and lithic lapilli-tuff are associated with this basalt.

The Rothea Formation, in ascending order, is divisible into the Sophia, Juvenile and Image Brook members. The Sophia Member is further divisible into a lower, unwelded, but compacted, pumiceous lapilli-tuff and two crystal tuff units. The upper unit invariably has 1% altered, platy biotite and accessory zircon. The Juvenile Member grades from nearly aphyric tuff at the base to crystal tuff at the top. Pyroxene(?) pseudomorphs typify this unit. The Image Brook Member, which has the least areal extent of the three members in the Rothea Formation, consists of a lower fine grained redbed unit and an upper lithic tuff unit.

The South Oromocto Andesite is composed of at least three flow units with the basal flow being the most areally extensive and the only one exhibiting porphyritic textures. Calcite veins and hematite bands near the top reflect degassing of the flow interior.

The Carrow Formation is predominantly a fining-upward redbed unit that grades from pebble- to cobble-conglomerate at the base to mudstone with intercalated calcrete at the top. Toward the southwest, the conglomerate contains abundant clasts of the Seelys (intracaldera sequence) and Rothea Formations, but to the northeast, metasedimentary clasts predominate. In the lower part of this formation, there is an unwelded, but highly compacted, pumiceous lapilli-tuff that contains abundant pumice fragments. Near the southwestern end of the Carrow outcrop belt, a basalt and basalt-clast mudflow occur near the top of the formation. A spore locality from the upper part of the Carrow Formation has yielded a precise Late Famennian age (McGregor and McCutcheon, 1988).

The Bailey Rock Rhyolite is a porphyritic lava and, like other lavas, is characterized by an absence of angular crystal fragments and pumice pseudomorphs. In places this rhyolite is intrusive into older rock units. It is unique because it crosses the boundary between the exocaldera and intracaldera sequences. A saprolite separates the Bailey Rock from the overlying late caldera-fill sequence.

The Intracaldera Sequence

The intracaldera rocks comprise, in strataigraphic order, the Scoullar Mountain Formation, Little Mount Pleasant Formation, Seelys Formation and McDougall Brook Porphyritic Microgranite. In addition, there are unnamed felsic dykes and one mafic dyke that intrude the Scoullar Mountain and Little Mount Pleasant formations, respectively.

The Scoullar Mountain Formation is characterized by sedimentary breccia (the argillite-talus breccia of previous authors) and interbedded andesitic lavas but felsic pyroclastic rocks are voluminous in places and one sandstone-conglomerate unit is present. The sedimentary breccia, though dominated by pebble- to boulder-size angular metasedimentary clasts, contains a few undeformed crystal tuff clasts that contain about 1% altered biotite. A pumiceous lapilli-tuff near the apparent top has about 1% amphibole and accessory apatite. The "brecciated metasediments" (Ruitenberg, 1967) at Mount Pleasant are part of this formation.

The Little Mount Pleasant Formation comprises two crystal tuff units and one flow-banded rhyolite. The tuff units are characterized by unflattened to weakly flattened, microcrystalline, recrystallized pumice fragments and chloritized amphibole with associated apatite. Phenocrysts inside these recrystallized pumice fragments are an order of magnitude larger than those outside, indicating that significant mechanical breakage of phenocrysts occurred during eruption. The "quartz-feldspar porphyry" (Ruitenberg, 1967) at surface on Mount Pleasant is correlative with the upper tuff unit; the quartz-feldspar at depth is probably correlative with crystal tuff of the Scoullar Mountain Formation.

The Seelys Formation, in ascending

stratigraphic order, consists of lithic tuffs and pumice-bearing lithic lapilli-tuffs; banded, pumiceous, crystal tuff; and densely welded crystal tuff. The basal unit contains clasts of Scoullar Mountain andesite and Little Mount Pleasant rhyolite. Quartz and feldspar phenocrysts increase in size and abundance from base to top in the upper unit. Platy biotite is virtually absent but metamict zircon is a common accessory in all three units. Rocks of this formation do not occur at Mount Pleasant as stated by previous authors.

The McDougall Brook Porphyritic Microgranite consists mostly of porphyritic monzogranite, but also includes a marginal or border phase feldspar ($\pm$ quartz) porphyry and minor equigranular to subporphyritic, fine-grained quartz monzonite. The groundmass grain size of the porphyry, the size and abundance of feldspar phenocrysts increase inward, away from the contact with country rocks. Chloritized amphibole with associated apatite is the main ferromagnesian mineral phase in all three units. Parts of the feldspar porphyry are hydrothermally altered, and a small hydrothermal breccia or diatreme cuts the microgranite. The "feldspar porphyry" at Mount Pleasant is correlative with the McDougall Brook rocks although Sinclair *et al.* (1988) infer that it is extrusive rather than intrusive at this locality.

The relative stratigraphic position of units in the exocaldera and intracaldera sequences is based on the following observations:

a) The upper part of the Sophia Member (Rothea Formation) consistently contains about 1% platy biotite pseudomorphs. The only intracaldera rocks with this much biotite are volcanic clasts within sedimentary breccia of, and a tuff unit near the apparent base of, the Scoullar Mountain Formation.

b) Andesitic rocks occur only in two units: the South Oromocto Andesite of the exocaldera sequence and the Scoullar Mountain Formation of the intracaldera sequence.

c) The Carrow Formation contains clasts from the Seelys Formation.

d) The Bailey Rock Rhyolite, which occurs in both sequences, intrudes and/or overlies the Carrow Formation but is intruded by, or grades into, the McDougall Brook Porphyritic Microgranite.

The Late Caldera-Fill Sequence

The late caldera-fill rocks include the Mount Pleasant Porphyry and its associated breccias, unnamed rapakivi dykes, the Big Scott Mountain Formation and the Kleef Formation. The first two units are separated spatially both from each other and from the other two formations. The ages of the late caldera-fill rocks are not firmly established; they are most likely Late Devonian but could range into the Mississippian.

The Mount Pleasant Porphyry is restricted to the Mount Pleasant area where it occurs as dykes and small, plug-like bodies associated with magmatic-hydrothermal breccias, as defined by Sillitoe (1985). The dykes commonly exhibit flow-banding, and crosscutting relationships between the dykes indicate multiple stages of intrusion. These dykes and plugs are the "banded porphyry" of Ruitenberg (1967). Two types of hydrothermal breccia are present: an older and more voluminous felsic phase, and a younger, chloritic phase (Kooiman et al., 1986). The porphyry, which grades into granitic rocks at depth, and associated breccias were emplaced at the <u>pre-existing</u> caldera margin.

Rapakivi dykes are composite bodies with leucocratic, porphyritic centres and melanocratic, sparsely porphyritic margins. The rocks are characterized by large phenocrysts of quartz and feldspar, up to 1 cm and 3 cm, respectively. In one locality, a porphyritic rhyolite lithologically similar to these dyke rocks is considered to be their extrusive equivalent.

The Big Scott Mountain Formation consists of, in ascending order, porphyritic to nearly aphyric rhyolite, lithic to lithic lapilli-tuff and crystal tuff. Most of the rhyolites are characterized by pyroxene(?) pseudomorphs. One of the rhyolite units appears to disconformably overlie McDougall Brook Porphyritic Microgranite. The lithic tuffs contain clasts that were derived from Seelys tuff, McDougall Brook Porphyritic Microgranite, and aphyric rhyolite of uncertain correlation.

Primary layering is discernible in the crystal tuff and is defined by slight differences in crystal size and abundance.

The Kleef Formation, from base to top, includes redbeds, porphyritic to glomeroporphyritic basalt and pumiceous, lithic tuff to lithic lapilli-tuff. Pebble- to cobble-conglomerate contains clasts of Scoullar Mountain and Seelys tuff, plus Bailey Rock and Big Scott Mountain rhyolite. The basalt is characterized by large plagioclase phenocrysts (up to 2 cm) and, near the top of the unit, some plagioclase glomerocrysts (up to several centimetres). The lithic tuffs are characterized by their reddish brown colour and abundant fossil-pumice.

The Piskahegan Group is disconformably overlain by post-volcanic redbeds of the Shin Formation. This unit postdates the middle Mississippian (Visean) Windsor Group and is gradationally overlain by the Pennsylvanian Petitcodiac (formerly Pictou) Group.

The tungsten-molybdenum and tin deposits are genetically related to granitic rocks that are co-magmatic with the Mount Pleasant Porphyry (Kooiman et al., 1986). Although this igneous activity was focused on the <u>dormant</u> caldera margin, the granite is not restricted to the margin because the associated gravity low (Williams, 1978) extends much farther west. The various granite phases identified by Kooiman et al. (1986) and the fluids that produced the mineral deposits were probably derived by *in situ* (i.e. no eruption) cooling of peraluminous anorogenic magma by double-diffusive fractional crystallization (McCutcheon, 1990). Mass balance calculations show that under the above conditions, small volumes (10-20 km^3) of magma with an initial composition like the Little Mount Pleasant Tuff could yield quantities of metal and fluid capable of producing the Mount Pleasant deposits.

Road Log: Mount Pleasant Caldera

Field stops in the Mount Pleasant caldera are shown as stops 1 to 5 in figure 1. Road log begins near Central Blissville at the intersection of highways 101 and 660, which can be reached by following Highway 7 from Fredericton to Geary and then Hwy 660 south. At the intersection of highways 101 and 660, turn west on Highway 101. Drive 1.0 km and turn left. Proceed 0.2 km and turn right on road to Mount Pleasant (gravel road no. 785). Drive 13.4 km to the large outcrop on the left side of the road.

STOP 1

Flow-banded rhylite at this location marks the base of the caldera fill sequence. The contact with the underlying porphyritic microgranite at the south end of the outcrop is concealed but the flow banding near the base dips gently northward. The rhyolite contains abundant spherulites and examples of the flow folds. Poorly preserved lithophysae can be seen in one place at the north end of the outcrop. Brecciated silica veins cut the outcrop in several places.

ROAD LOG Continue southward for 6.8 km to the intersection with lumber road no. 61. Turn left and proceed 0.4 km and turn right. Drive 1.8 km and park. Walk to the top of the knoll on the right side of the road and then down over the side.

STOP 2

Seelys Porphyry: crude columnar jointing and excellent eutaxitic foliation perpendicular to it can be seen in several outcrops at this location. The reddish streaks are flattened and stretched pumice; these are best observed in outcrops near the bottom of the hill.

Road log Return to the main road and turn left (south). Drive 3.6 km to the outcrops by the Piskahegan Stream.

STOP 3

North of the stream on the left side of the road are outcrops of pink feldspar porphyry, the border phase of the porphyritic microgranite. Note the glassy nature of the groundmass.

South of the stream, Little Mount Pleasant Tuff is exposed on the right side of the road. Effects of hydrothermal alteration and in situ brecciation can be seen in places.

Road log Continue south for 2.1 km to the intersection of lumber road on the left. Turn left, drive 0.3 km and stop.

STOP 4

Silicified breccia is exposed on the north side of the road. Recognizable fragments in the breccia include Little Mount Pleasant Tuff and feldspar porphyry; many are unidentifiable. Some fragments exhibit evidence of rounding but most are angular. The breccia is part of a small enclave within the porphyritic microgranite and is interpreted as a hydrothermal breccia, i.e. nonvolcanic.

Porphyritic microgranite is exposed about 100 m farther east on the north side of the road. At this point it is reddish; 200 m farther along on the south side of the road the microgranite is greenish and contains inclusions of older granitoid rocks. 300 m farther along on the north side of the road, rocks exposed are transitional between microgranite and feldspar porphyry.

Road log Return to the main road and continue sout for 6.9 km. At the Y-intersection, turn right and drive 2.1 km to the tributary of Hatch Brook.

STOP 5

Little Mount Pleasant Tuff; quartz-feldspar crystal tuff with moderately well developed eutaxitic foliation is exposed in the stream bed on the right (east) side of the road. The foliation is outlined by dark green fiamme or extremely flattened pumice fragments; this is one of few outcrops of this unit where eutaxitic foliation can be seen. This unit is similar to the quartz-feldspar porphyry unit which is one of the host rocks for the Mount Pleasant tungsten-molybdenum and tin deposits.

THE MOUNT PLEASANT TUNGSTEN-MOLYBDENUM AND TIN DEPOSITS[1]

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Host Rocks

As noted by McCutcheon (this volume), the Mount Pleasant deposits are associated with hydrothermal breccias and intrusive rocks that cut the intracaldera sequence. Various aspects of the geology at Mount Pleasant have been described by Ruitenberg (1963, 1967), Dagger (1972), Parrish and Tully (1978), Kooiman et al. (1986) and Sinclair et al. (1988).

The four main rock units exposed at surface (Fig. 1) were originally recognized by Ruitenberg (1967), although their correlation with the caldera stratigraphy was established by McCutcheon (1987). Ruitenberg's "brecciated metasediments" (unit SBx, Fig. 1) are correlative with sedimentary breccias of the Scoullar Mountain Formation; his "quartz-feldspar porphyry" (QFP) is equivalent to Little Mount Pleasant tuff; his "feldspar porphyry" (FP) is the extrusive(?) equivalent of McDougall Brook Porphyritic Microgranite; and his "banded porphyry" (GP/GRII) is Mount Pleasant Porphyry. The first three units belong to the intracaldera sequence; the last one is a late caldera-fill unit.

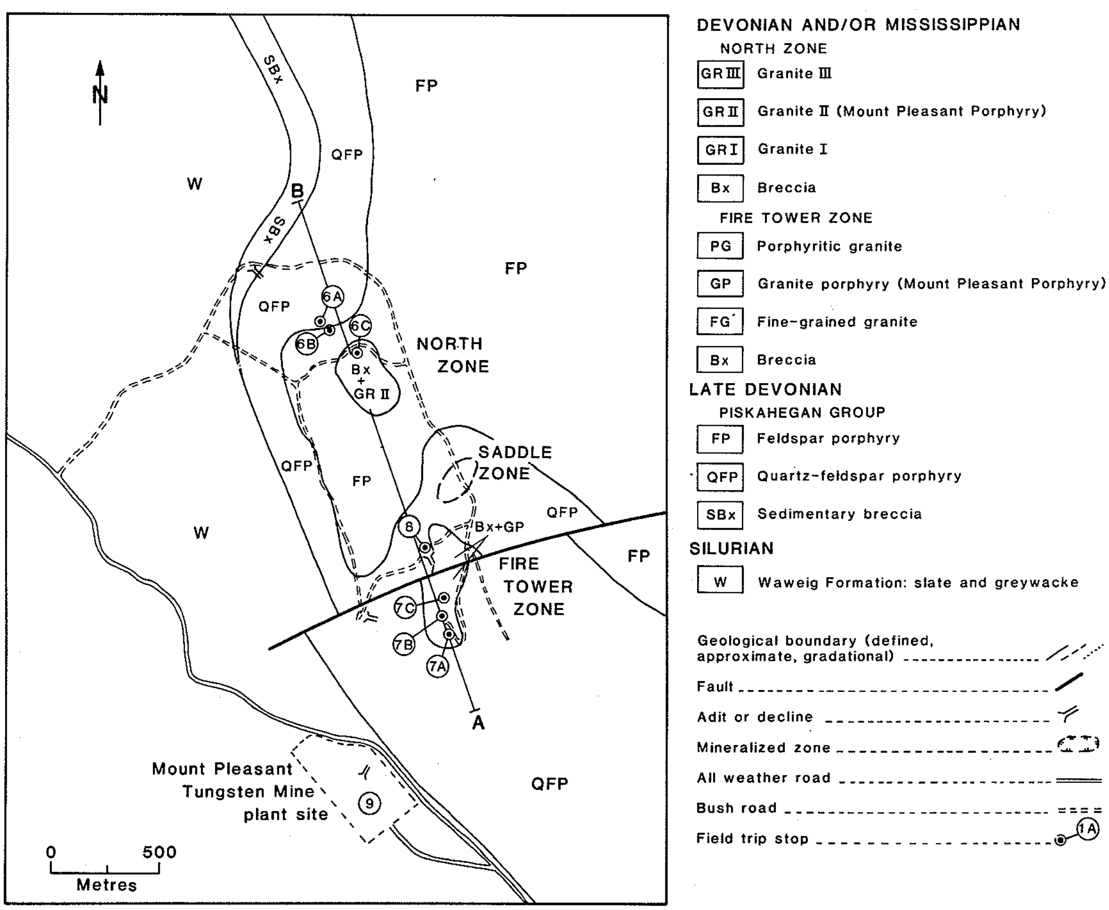

Figure 1. Geological map of the Mount Pleasant area (modified from Ruitenberg, 1967 and Sinclair et al., 1988).

[1]This paper was published previously in a Geological Association of Canada guidebook by Procyshyn et at. (1989) and is reprinted here with minor revisions.

Two additional felsic pyroclastic (QFP) units that are separated by a layer of sedimentary breccia occur at depth. This breccia, similar to the one exposed at surface, is also likely correlative with the Scoullar Mountain Formation. Granitic rocks intrude and underlie these units.

The intracaldera rocks have been highly brecciated, altered and mineralized in two areas at Mount Pleasant designated the North Zone and the Fire Tower Zone (Fig. 1). In both areas, crosscutting breccias and associated intrusive rocks form irregular, roughly vertical, pipelike complexes that were centres of subvolcanic intrusive and related hydrothermal activity. Smaller intrusive centres with little or no surface expression occur in the Saddle Zone (Fig. 1) between the North Zone and the Fire Tower Zone and at Hornet Hill. The breccias are composed of multiple phases which range from matrix-supported breccias with rounded fragments to clast-supported breccias with mainly angular fragments. Both fragments and matrix material have been altered extensively and in most places the textures and lithologies of the fragments are difficult to identify. Two types of breccia are recognized. One is characterized by pervasive, siliceous alteration consisting of fine grained quartz and topaz whereas the other is typified by chlorite and biotite alteration. Formation of the breccias and associated hydrothermal alteration likely were formed by repeated episodes of explosive brecciation related to crystallization of fluid-saturated granitic magma (Kooiman et al., 1986; Sinclair et al., 1988).

Distribution of, and contact relationships between, the various granitic units at Mount Pleasant (Fig. 2) have been determined mainly from drill core and exposures in underground workings. Units in the North Zone, from oldest to youngest, have been designated Granite I, Granite II and Granite III; units in the Fire Tower Zone are referred to as fine grained granite, granite porphyry and porphyritic granite.

Granite I occurs as irregular bodies closely associated with the North Zone breccias. Its contacts with the breccias are commonly gradational and fragments of Granite I are abundant locally within the breccias. Granite I is typically fine grained and equigranular in relatively unaltered specimens. However, in most areas textural features of Granite I have been obscured by pervasive chloritic and/or silicic alteration. Porphyry tungsten-molybdenum deposits in the North Zone appear to be related to Granite I.

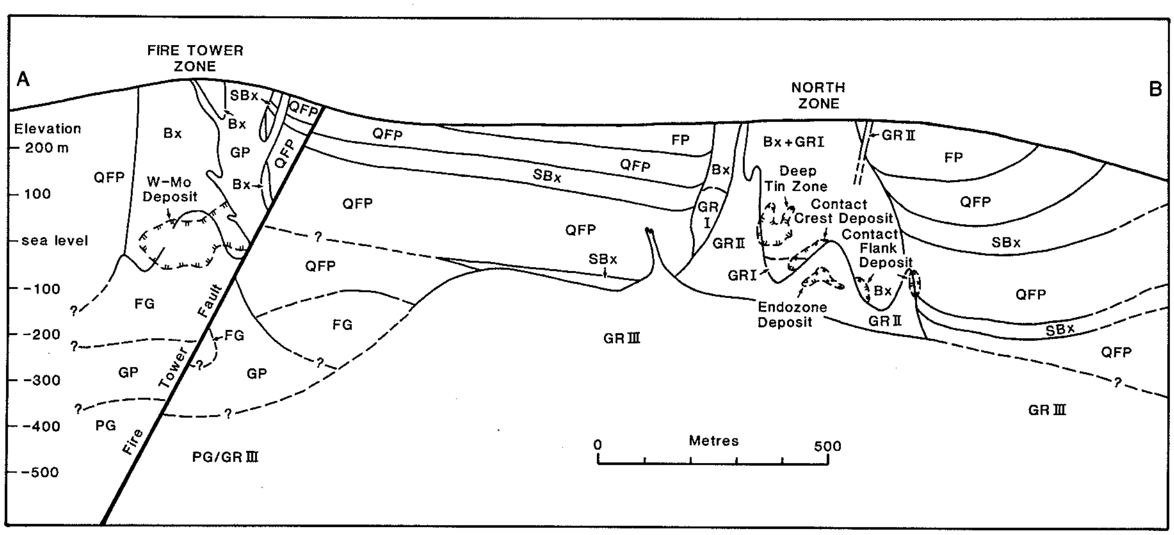

Figure 2. Cross-section A-B through the Mount Pleasant are (modified from Sinclair et al., 1988).

The main body of Granite II occurs at the base of the North Zone complex (Fig. 2) although in places dyke-like bodies of Granite II have intruded the overlying breccias and associated host rocks. Porphyry dykes that outcrop at surface are probably related to Granite II.

Granite II varies from aplitic to porphyritic in texture. Parts of Granite II also contain abundant miarolitic cavities and comb quartz layers. The comb quartz layers consist of parallel to subparallel layers in which quartz crystals are oriented approximately perpendicular to the planes of layering (cf. Kirkham and Sinclair, 1988). They are one of a family of unidirectional solidification textures (USTs) that are associated with fluid saturated and/or undercooled magmas (cf. Shannon et al., 1982).

Tin deposits in the North Zone are associated mainly with Granite II. The Endozone, Contact Crest and Contact Flank deposits, for example, occur either within or at the contact of Granite II (Fig. 2). The Deep Tin Zone is also likely related to Granite II, although it is not as closely associated spatially. Tin zones at surface in the North Zone are associated with porphyry dykes that are likely offshoots of Granite II.

Granite III forms a large body that underlies both the North and Fire Tower zones. In the North Zone, contact relationships show clearly that Granite III has intruded granites I and II. The contacts between Granite III and the other granites are commonly sharp and in many places are marked by thin (0.5 to 2 cm wide) layers of USTs, mainly K-feldspar, in Granite III. Chemical data indicate that Granite III is identical to porphyritic granite of the Fire Tower Zone.

Granite III is typically fine- to medium-grained and equigranular although porphyritic and pegmatitic varieties are also present. Miarolitic cavities filled with very fine grained sericite are locally abundant. Granite III is generally less fractured and altered than granites I and II, and only a few isolated tin zones have been found within it.

Fine-grained granite underlies the breccia complex in the Fire Tower Zone (Fig. 2). Directly beneath the breccia this granite is highly brecciated and cut by a stockwork of molybdenum- and tungsten-bearing fractures and quartz veinlets. Silicic and chloritic alteration occur along the selvages of the fractures and quartz veinlets but in highly fractured rocks, this alteration is pervasive. The density of fracturing and intensity of alteration and mineralization decrease gradually with depth. Relatively unaltered fine-grained granite generally has an aplitic texture, although phenocrysts of quartz and K-feldspar are present in places. Porphyry tungsten-molybdenum deposits in the Fire Tower Zone are related to the fine-grained granite; in this regard the fine-grained granite is comparable to Granite I of the North Zone.

Granite porphyry (Mount Pleasant Porphyry) is irregularly distributed in the Fire Tower Zone. At depth, fine grained granite grades into granite porphyry. At higher levels granite porphyry has intruded fine grained granite, Fire Tower breccia and quartz-feldspar porphyry and forms dyke-like bodies that pinch and swell with numerous offshoots or extensions (Fig. 2). Contacts of the granite porphyry have chilled margins and commonly display flow layering and small xenoliths of adjacent wall rock.

Relatively unaltered granite porphyry is, in places, similar in texture to fine-grained granite but is more typically characterized by quartz and K-feldspar phenocrysts. Comb quartz layers, similar to those in Granite II in the North Zone, occur locally. Much of the granite porphyry has been pervasively sericitized and/or chloritized; chloritized granite porphyry has the appearance of a quartz porphyry. Tin-bearing veins and replacement zones in the Fire Tower Zone appear to be spatially associated with granite porphyry.

With increasing depth, granite porphyry grades into a fine- to medium-grained, equigranular to porphyritic granite. The distribution of this granite in the Fire Tower Zone is poorly known although it appears to be directly related to Granite III in the North Zone. Like Granite III, porphyritic granite contains only a few poorly mineralized zones.

Age of the Mount Pleasant Intrusions

The absolute age of the Mount Pleasant intrusions is uncertain. K-Ar and Rb-Sr studies (Kooiman et al. 1986) indicated a Late Mississippian age of 340 to 330 Ma. On the other hand, preliminary $^{40}Ar/^{39}Ar$ data suggest a Late Devonian-Early Mississippian age of about 360 Ma (D.A. Archibald, personal communication, 1985), essentially the same age as the Piskahegan Group rocks which the granite have intruded. A K-Ar date of 361 ± 9 Ma recently obtained at GSC from biotite hornfels in sedimentary breccia that is underlain by Granite III appears to confirm this Late Devonian-Early Mississippian age. More accurate and precise dating of the granitic rocks by U-Pb analysis of monazite might resolve this discrepancy. Rb-Sr isotopic studies of fluorite associated with the various granitic intrusions and associated deposits, currently underway by R.P. Taylor (Carleton University), may also contribute to a resolution of the age of the Mount Pleasant intrusions.

Geochemistry of the Mount Pleasant Intrusions

The chemical compositions of five representative samples of Mount Pleasant granitic rocks are given in Table 1. The samples are typical of the least altered granites; no composition of Granite I is given because of the lack of relatively unaltered samples.

The intrusive rocks are high silica (74-77% SiO_2), fluorine-rich (>3000 ppm F) granites. Compared to average granites, they are Ca-poor and alkali-rich, with more K_2O than Na_2O. They have elevated contents of Li, Rb, Y, Nb, Cs, Ga, Sn and W, and low concentrations of MgO, P_2O_5, TiO_2, Sr and Ba. Although the composition of the different intrusive phases is similar with regard to many of the elements, the youngest phases, porphyritic granite and Granite III, have significantly higher contents of Rb and Li than the other phases and appear to be the most chemically evolved. The chemical data, and the local occurrence of primary topaz in an aplitic contact phase of Granite II and as inclusions in euhedral quartz crystals in Granite III, indicate that the Mount Pleasant intrusions are compositionally analogous to topaz granites (cf. Manning, 1988).

Tungsten-Molybdenum Deposits

Tungsten-molybdenum deposits in both the North Zone and the Fire Tower Zone are large, low-grade, deposits. Reserves in the Fire Tower Zone prior to mining totalled 22.5 million tonnes grading 0.21% W, 0.10% Mo and 0.08% Bi; approximately 11 million tonnes of similar grade material are present in the North Zone (Parrish and Tully, 1978). Included in these reserves was a higher grade deposit in the Fire Tower Zone containing 9.4 million tonnes grading 0.39% WO_3 and 0.20% MoS_2. During the two years of mining this deposit from 1983 to 1985, the Mount Pleasant Tungsten Mine produced more than 2000 tonnes of concentrate grading 70% WO_3 from about one million tonnes of ore.

The tungsten-molybdenum deposits are hosted mainly by breccia and, to a lesser extent, by associated country rocks; in the Fire Tower Zone, they are also hosted by fine-grained granite that underlies the breccia. The deposits consist of mineralized fractures, quartz veinlets and disseminations in breccia matrix. Wolframite and molybdenite are the principal ore minerals; minor amounts of bismuth and bismuthinite are also present. Quartz, topaz, fluorite, arsenopyrite and loellingite are the principal gangue minerals.

Alteration associated with the tungsten-molybdenum deposits includes several different types. Intense and pervasive silicic or greisen-type alteration occurs within and above the higher grade tungsten-molybdenum zones. This type of alteration is characterized by the complete or nearly complete replacement of host rocks by quartz, topaz and fluorite. This alteration grades outward to a less intense silicic alteration that is limited mainly to narrow selvages on mineralized fractures and quartz veinlets. Quartz, biotite, chlorite and minor amounts of topaz are the principal minerals of this alteration stage which extends laterally up to 100 m beyond the higher grade tungsten-molybdenum zones. Propylitic alteration consisting of chlorite and sericite surrounds the silicic alteration and extends for more than 1000 m before grading into relatively unaltered rock.

Table 1. Chemical composition of selected samples of Mount Pleasant granitic rocks.

	84-4	82-33	84-18	87-11	82-50
SiO_2	76.8	74.2	76.0	75.9	75.3
TiO_2	0.05	0.04	0.06	0.04	0.03
Al_2O_3	12.9	13.0	13.3	12.9	13.5
Fe_2O_3	0.1	0.7	0.0	0.0	0.3
FeO	1.0	0.5	1.2	1.5	1.0
MnO	0.05	0.05	0.08	0.03	0.08
MgO	0.0	0.19	0.13	0.02	0.09
CaO	0.6	0.86	0.87	0.70	0.52
Na_2O	3.3	3.0	3.1	3.3	3.1
K_2O	4.47	4.85	4.58	4.92	4.86
H_2O	0.7	0.8	0.6	0.8	0.5
CO_2	0.1	0.1	0.1	0.1	0.1
P_2O_5	0.03	0.03	0.03	0.0	0.01
S	0.07	0.0	0.02	0.0	0.02
F	0.35	0.85	0.99	0.55	0.78
$-O=F_2+S$	0.18	0.36	0.43	0.23	0.34
Total	100.3	98.8	100.6	100.5	100.4
Ba	20	80	30	86	90
Cs	9	13	21	7	20
Rb	670	888	1200	823	1136
Sr	12	13	19	10	20
Li	45	157	610	220	349
Nb	79	90	68	59	85
La	36	35	44	94	42
Ce	103	89	130	147	107
Y	220	157	120	164	156
Zr	160	146	99	94	143
Ga	29	-	30	26	29
W	9	10	14	12	8
Sn	<3	25	60	22	100
Mo	10	<1	10	7	<1
Cu	3	9	2	10	6
Pb	45	26	37	64	66
Zn	54	75	32	68	127

Major elements, Ba, Rb, Sr, Nb, Y by XRF analysis; FeO by wet chemical analysis; H_2O, CO_2, S by infrared spectrometric analysis; F by selective ion electrode analysis; Li, Cu, Pb, Zn, W, Ga by ICP emission analysis; Analytical Chemistry Section, Geological Survey of Canada. Sn analyses by emission spectrometric analysis, X-Ray Assay Laboratories, Don Mills.

Samples:

84-4: Fine grained granite, Fire Tower Zone.
82-33: Granite porphyry, Fire Tower Zone.
84-18: Porphyritic granite, Fire Tower Zone.
87-11: Granite II, North Zone.

Tin deposits

Tin-bearing, sulphide-rich polymetallic vein and replacement deposits are superimposed on the tungsten-molybdenum deposits in both the North and Fire Tower zones. In these deposits, very fine grained cassiterite and stannite are associated with chlorite, fluorite and complex assemblages of sulphides and sulpharsenides including arsenopyrite, loellingite, sphalerite, chalcopyrite, galena, pyrite, marcasite, molybdenite, tennantite, bornite, bismuthinite, wittichenite and roquesite (Petruk, 1973a). These deposits are generally small; proven reserves are approximately 250,000 tonnes grading 0.6% Sn, 2.3% Zn, 0.30% Cu and 0.36% Pb (Mount Pleasant Mines Limited, Annual Report for 1973 as reported in Mulligan, 1975). Based on a review of Mount Pleasant Mines Ltd. data, reserves in the No. 7 lode in the Fire Tower Zone are estimated to be about 15,000 tonnes grading 0.5% Sn, 5.6% Zn, 1.2% Pb and 0.7% Cu.

Most of the potentially economic tin deposits occur in the North Zone at depth of 200 to 400 m below surface. They include the Deep Tin Zone, Contact Crest, Contact Flank and Endozone deposits (Fig. 2). The Deep Tin Zone is a relatively large, irregular deposit that consists of fracture-controlled and disseminated cassiterite in silicified and chloritized breccia and Granite I. Other minerals associated with cassiterite include arsenopyrite, sphalerite, chalcopyrite and galena. Reserves in the Deep Tin Zone are approximately 2.14 million tonnes grading 0.45% Sn, 0.06% W, 0.03% Mo and 0.80% Zn (the W and Mo grades are probably due to the superposition of the Deep Tin Zone on a tungsten-molybdenum zone; Sinclair et al., 1988).

The Contact Crest and Contact Flank deposits occur mainly in breccia or other associated host rocks at the upper contact or along the sides of Granite II. The Endozone deposit, on the other hand, occurs mainly within Granite II. In these deposits, cassiterite occurs as finely disseminated grains and as fine- to medium-sized grains in veins or veinlets and along fractures. Associated minerals include arsenopyrite, sphalerite, chalcopyrite, pyrite and pyrrhotite. Chlorite, fluorite, quartz, topaz and sericite are the main alteration minerals. Crosscutting relationships indicate that as many as 6 stages of alteration and mineralization may be present. Total geological reserves in the Deep Tin Zone, Contact Crest, Contact Flank and Endozone deposits have been estimated at 5.1 million tonnes averaging 0.79% Sn (<u>The Northern Miner</u>, 06 November 1989, p. A2).

Recent exploration of the newly-discovered Saddle Zone area indicated in figure 1 has indicated potential for an important new tin zone associated with a previously unknown cupola of Granite II. One drill hole in this area intersected 16 m grading 1.31% Sn and included a section of 5.1 m that assayed 3.97% Sn (<u>The Northern Miner</u>, 22 January 1990, p. 3).

Road Log Mount Pleasant Tungsten-molybdenum and tin deposits

Field stops for this excursion are shown as stops 6 to 9 in figure 1 (previous section).

Road log Continue from STOP 5 in the Mount Pleasant caldera for 4.0 km, past the Mount Pleasant Project site (former Mount Pleasant Tungsten Mine). Turn right on bush road and drive 1.0 km. Turn right and dirve 0.5 km. Bear left at Y-intersection and proceed 40.4 km to area fo large trenches. Park. Walk 175 m to the northwest along the cleared area.

STOP 6A

Quartz-feldspar porphyry; the quartz-feldspar porphyry at this location is slightly chloritized but can be recognized by quartz and feldspar phenocrysts typically 2 to 3 mm in size. Fractures and quartz veinlets in the quartz-feldspar porphyry locally contain sulphide minerals. Walk 15 m back toward the road.

STOP 6B

Feldspar porphyry; fractured and altered. The feldspar porphyry is distinguished from quartz feldspar porphyry in having feldspar phenocrysts up to 1 cm in size and generally less than 5% quartz phenocrysts. The contact between the feldspar porphyry and the quartz-feldspar porphyry is marked by the local change in slope. Walk back to the trenched area on the south side of the road.

STOP 6C

North Zone; Fine-grained tin materials (cassiterite and stannite) and base metal sulphides occur locally in fractured, brecciated and intensely silicified feldspar porphyry intruded by dykes of felsic porphyry.

Road log Continue on bush road 0.3 km to T-intersection and turn right. Proceed 1.0 km, bearing to the right at the Y-intersection, and continue for another 0.2 km. Park and continue by foot along the road for 300 m to the top of the hill near the old fire tower.

STOP 7A

Silicified breccia intruded by at least two generations of Mount Pleasant Porphyry dykes can be seen in outcrops to the southwest of the fire tower.

STOP 7B

Highly silicified felsic breccia, breccia dykes and chloritized and silicified Mount Pleasant Porphyry (granite porphyry) are exposed in the large area of outcrop northwest of the fire tower.

STOP 7C

Fluorite, molybdenite and wolframite occur in fractures and quartz veinlets in silicified felsic breccia exposed in the large trench at the north end of the large area of outcrop. The felsic breccia is intruded by a dyke of Mount Pleasant Porphyry that locally contains comb quartz layers.

Road log Return 0.4 km back along the bush road to the intersection with a bush road on the left (west) side. Park and walk 240 m west along the bush road to the site of the 900 adit and dump.

STOP 8

900 adit; Mount Pleasant Porphyry is exposed in road on east side of the adit, fractured and altered quartz-feldspar porphyry occurs on the west side of the adit. The adit (no longer accessible) was developed to explore a small tin-bearing sulphide replacement deposit (the No. 7 lode); samples of massive sulphides including sphalerite, galena, chalcopyrite, cassiterite and stannite may be found on the dump on the north side of the road. Roquesite ($CuInS_2$) has been identified in polished sections of samples from this area (Petruk, 1973a).

Road log Return to Lac-Billiton Tin Project site.

STOP 9

Examine drill core and rock specimens from the Mount Pleasant tungsten-molybdenum and tin deposits. Cassiterite-bearing samples from the Endozone deposit may be found in the parking lot south of the main gate. Metasedimentary rocks of the Lower Devonian Waweig Formation that form the walls of the Mount Pleasant caldera are exposed in several places on the mine site.

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THE EAST KEMPTVILLE POLYMETALLIC TIN DOMAIN

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INTRODUCTION

The East Kemptville Polymetallic Tin Domain was first proposed by Chatterjee in 1983 to encompass a linear belt of granophile and chalcophile mineral deposits within, and extending from, the south-western extension of the South Mountain Batholith (Fig. 1). A voluminous amount of literature is available or in preparation on this domain and only the salient points can be given here. The reader is referred to the extensive bibliography given at the end of this description if more detailed information is required.

The East Kemptville Polymetalic Tin Domain is located within the Meguma Terrane, the easternmost of several terranes that collectively constitute the Canadian Appalachians (Williams and Hatcher, 1983, Fig. 1 of guidebook introduction). This terrane was accreted during the Late Devonian Acadian Orogeny via emplacement along the east-west Cobequid-Chedabucto Fault System (Webb, 1969; Keppie, 1982; Mawer and White, 1986). Continued transpression of this terrane is documented by the presence of numerous northeast- southwest to east-west trending shear zones throughout the Meguma Terrane which have been variably constrained to have been active at ca. 370-300 Ma (Keppie and Dallmeyer, 1987; Dallmeyer and Keppie, 1987; Reynolds et al, 1987; Hill, 1988; Muecke et al., 1988). Recently Giles (1985) has proposed the presence of a major transcurrent fault zone within the Meguma Terrane which he named the Tobiatic Shear Zone. A proposed post-Visean sinistral displacement along this zone of 110 km allows reconstruction of the Meguma Terrane as shown in figure 2 (Giles, 1985). Although the sinistral displacement along this zone has not been verified by field mapping a number of local shear and breccia zones coinciding with this zone have been noted (see Giles, 1985). In the East Kemptville area the informally named Rushmere Lake Shear Zone (Fig. 3) occurs along the southeastern extremity of the proposed Tobiatic Shear Zone (Smith, 1985). Parallel to the Tobiatic Shear Zone is the proposed East Kemptville-East Dalhousie Shear Zone which has its expression in the East Kemptville area along the northern granite-Meguma contact incorporating the East Kemptville Tin deposit and Meguma-hosted tin occurrences (Fig. 3; Kontak, 1987a,1987b; O'Reilly, 1988).

The Meguma Terrane is underlain by inferred Precambrian ortho- and paragneiss based on recent field and geochemical investigations (Giles and Chatterjee, 1986,1987; Chatterjee and Giles, 1988; Clarke and Chatterjee, 1988). Overlying the inferred basement rocks with presumed unconformity are Lower Paleozoic meta-sedimentary and metavolcanic rocks (e.g. Meguma Group, White Rock Formation). These rocks were deformed and metamorphosed during the Mid to Late Devonian Acadian Orogeny (Keppie, 1982; Muecke, 1984; O'Brien, 1985).

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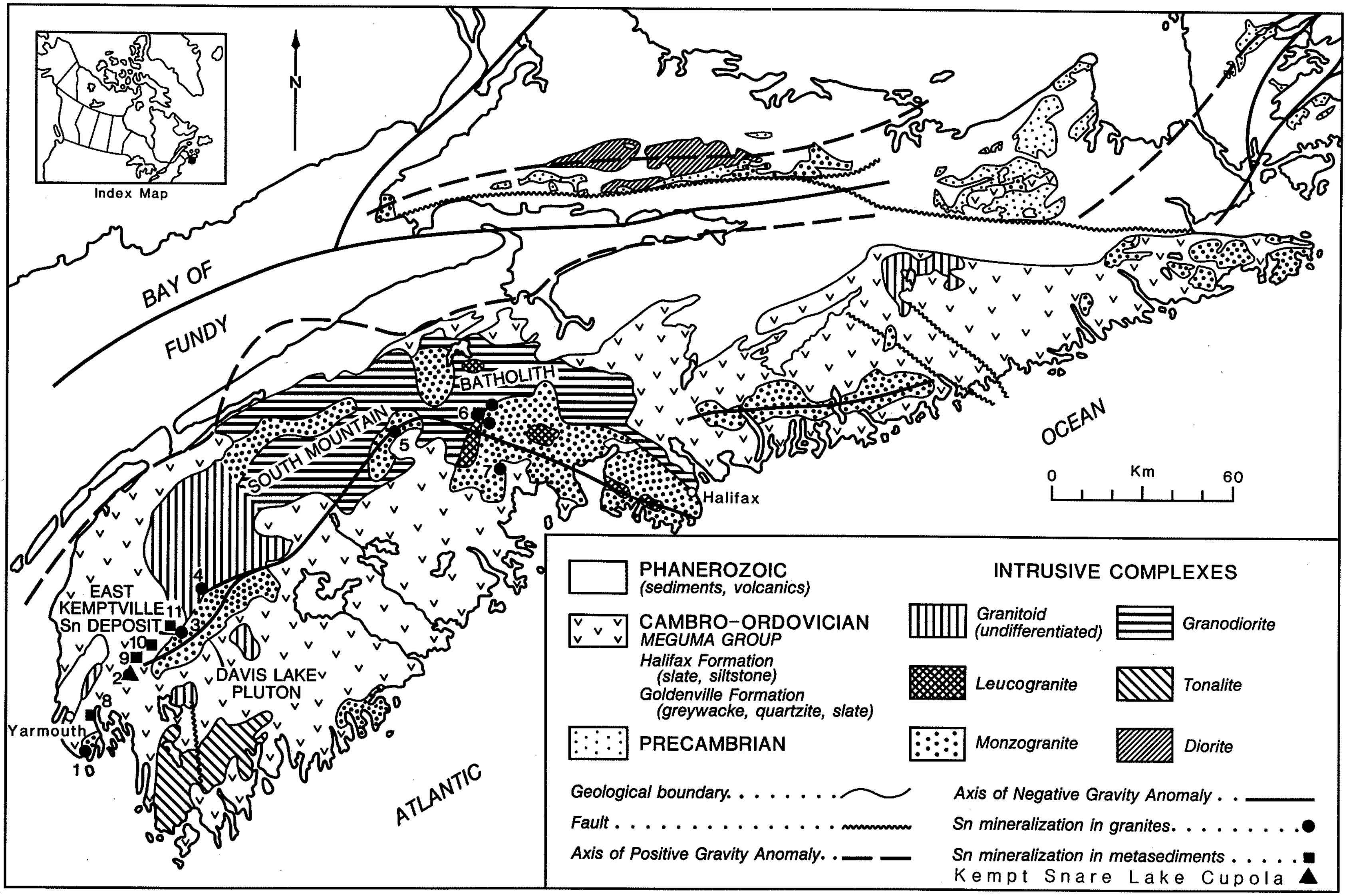


Figure 1. Geology of Meguma Terrane in Nova Scotia showing location of Davis Lake Pluton, East Kemptville Tin deposit and other tin occurrences in granites and metasediments. Tin mineralization in granite (1. Wedgeport Pluton, 3. East Kemptville Tin deposit, 4. Oakland Lake, 5. East Dalhousie, 6. New Ross, 7. Long Lake). Tin in metasediments (8. Dominique-Plymouth, 9. Pearl Lake, 10. Gardners Meadow, 11. Duck Pond). Kempt Snare Lake (W-Pb-Zn-As-Cu-Ag) cupola 2., (after Boyle, 1988).

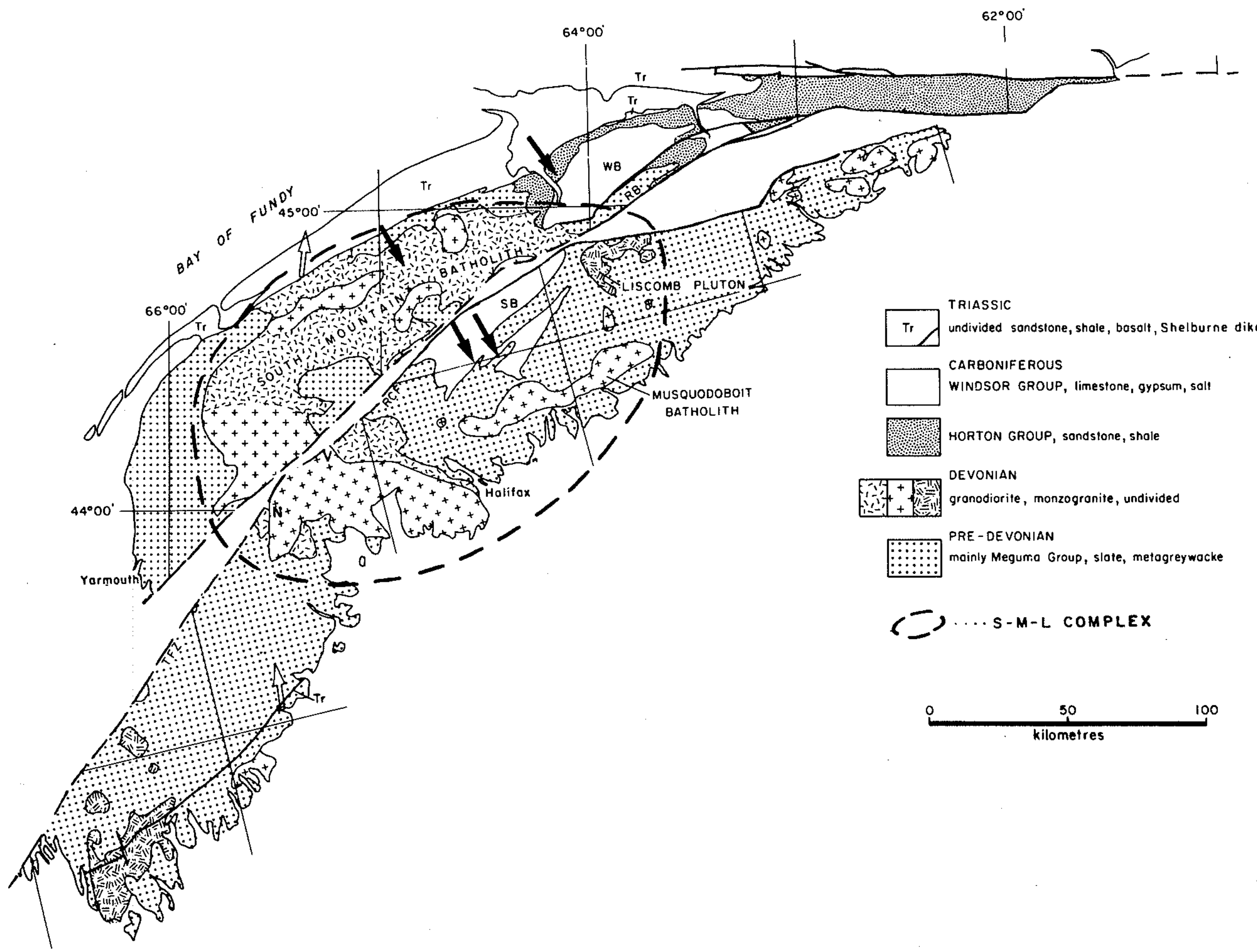


Figure 2. Reconstructed geology of southern Nova Scotia removing 110 km of sinistral offset. Note the realignment of structural and stratigraphic features and the restored configuration of major plutonic bodies as an elliptical plutonic complex (after Giles, 1985).

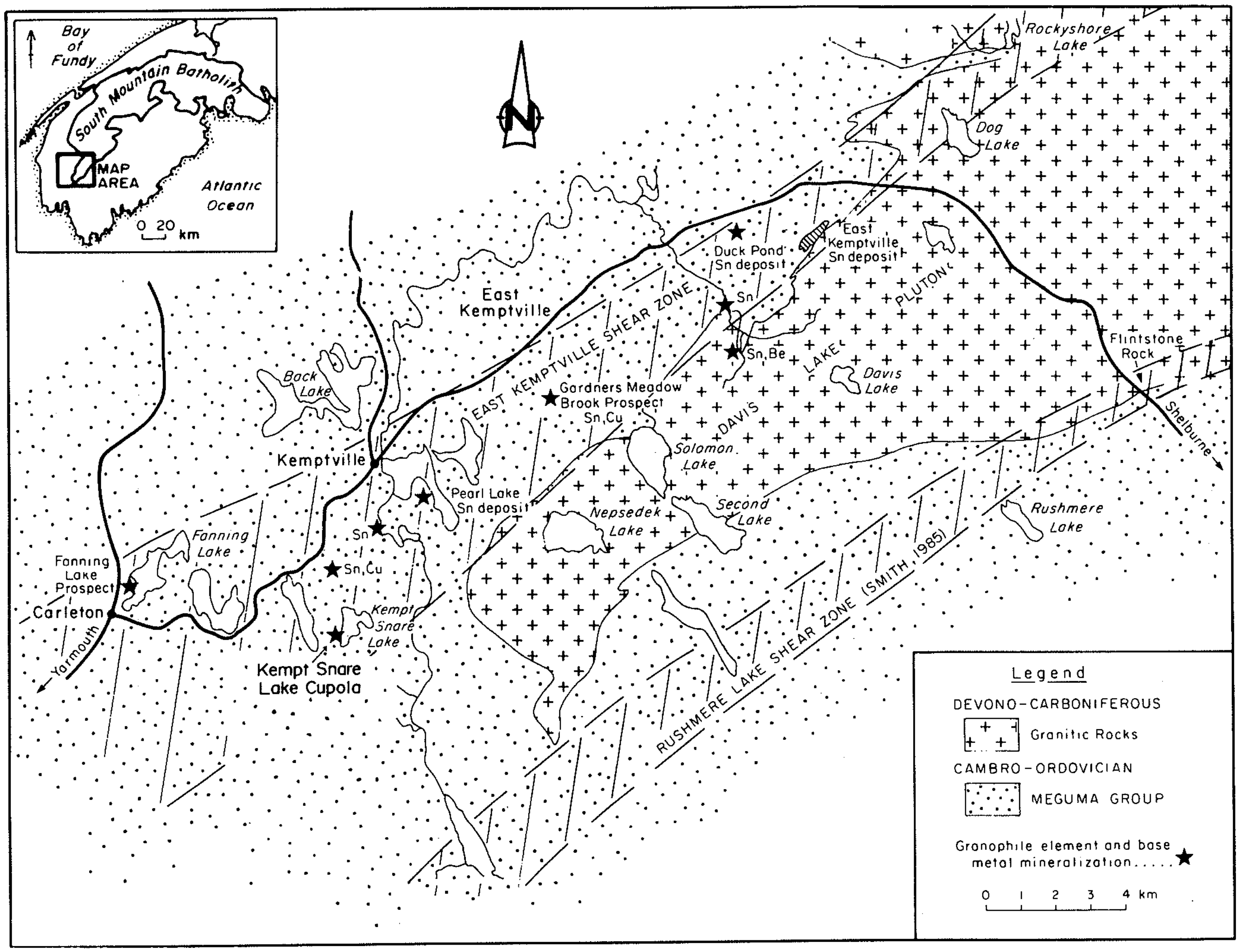


Figure 3. General geology of East Kemptville region showing locations of mineralized zones and extent of East Kemptville and Rushmere Lake shear zones (after Soehl et al., 1987).

The Meguma Terrane is composed of a 14-km-thick Cambro- Ordovician sequence of metawacke and meta-argillite intruded by peraluminous granites. Although whole-rock Rb-Sr data from the metasedimentary rocks do not yield isochrons (Clarke and Halliday 1980; Lambert et al. 1984), the average $^{87}Sr/^{86}Sr$ ratios at 330 Ma are 0.71282 (5 metawacke samples) and 0.71733 (6 meta-argillite samples) (data from Clarke and Halliday 1980). The post-tectonic South Mountain Batholith, an amphibole-free, tourmaline-bearing, comagmatic suite of biotite granodiorite, monzogranite and "porphyry", intruded between 372 and 361 Ma; initial $^{87}Sr/^{86}Sr$ ratios range from 0.7076 to 0.7102 (Clarke and Halliday 1980). Older, possibly syn- or pre-kinematic granite and diorite occur south of the Davis Lake Complex (Fig. 1). K-Ar, $^{40}Ar/^{39}Ar$ and fission-track dates summarized by Elias (1986), Dallmeyer and Keppie (1987) and Reynolds et al. (1987) indicate that granitic rocks of the Meguma Terrane range in age from 370-240 Ma and have had a complex tectonothermal history.

In the Davis Lake area, the Meguma Group consists of metawacke of the Goldenville Formation, a transitional meta-argillite - metawacke unit, and the Halifax Formation slate. These units are folded into northeast-trending, north-plunging folds. Tin and base-metal showings, as well as geophysical and geochemical anomalies, are aligned within the transition unit. This alignment, gravity data (Egner 1987) and the nature of the granite-metawacke contact near the deposit (Richardson 1988b) imply that the sub- surface topography of the contact consists of two undulating ridges. These ridges are sub parallel to the exposed granite- metawacke contact and the regional fold axes. One ridge occurs below the transition unit; the other at the present contact.

To the north and west, the Davis Lake Complex intrudes Goldenville metawacke and appears to intrude granodiorite of the Devonian South Mountain batholith. To the south and east, the Davis Lake Complex is in either fault or intrusive contact with metawacke: outcrop and drill-core are lacking. A biotite-bearing thermal aureole overprints greenschist-facies metawacke to the west. However, on the east side of the complex, the rocks were subjected to amphibolite to cordierite facies regional metamorphism.

The granite-metasedimentary contact in the East Kemptville area is undulose and sharp; assimilation and stopping of Meguma material is not evident. Granitic dyke rocks in the Meguma are rare and when present penetrate only a short distance into the metasediments.

EXPLORATION HISTORY

Tin mineralization, in the form of small granite-hosted greisen and pegmatite deposits, was first discovered in the New Ross area of Nova Scotia in the early 1900's (Fig. 1, No. 6). Although it has been long recognized that the South Mountain Batholith has similar tin-bearing characteristics to the tin-bearing Hercynian intrusions of Europe, it was not until 1975 that systematic exploration for tin in this batholith was begun. The first occurrence of greisen-style tin mineralization in the East Kemptville area was discovered at Pinkneys Point in the Wedgeport Pluton (No. 1, Fig. 1; McAuslan et al,1980). Later, in this same area, Meguma-hosted tin mineralization was discovered in glacial erratics. Following these finds, Shell Canada Resources in conjunction with the Millmor Syndicate carried out an extensive exploration program for tin in the area from 1975 to 1978. Application of airborne EM , ground magnetometer, boulder tracing and till geochemistry resulted in the discovery of the East Kemptville, Dominique and Duck Pond deposits (No. 3, 8 and 11, Fig. 1). Later till geochemical surveys in the area by Esso Minerals Canada and Falconbridge Limited delineated the Pearl Lake and Kempt Snare Lake deposits respectively (No. 2 and 9, Fig. 1).

A detailed till geochemical survey by Shell Canada Resources over the Davis Lake Complex (approx. 3 samples/km^2) outlined a large dispersion train with a width of 3 km extending about 6-7 km in a southeasterly direction from the East Kemptville and Duck Pond deposits (Boyle, 1988). Tungsten which is concentrated in the East Kemptville deposit (58 ppm in ore compared to 6 ppm in biotite granite), also shows a dispersal pattern similar to tin (Rogers and Garrett, 1987).

Analysis of airborne radiometric data over this area shows it to be an effective method for: A). outlining hydrothermal and magmatic differentiation trends related to granite-hosted mineralization, B). mapping the granite-Meguma contact where drilling information is lacking, and C). delineating the effects of glacial transport (Boyle, 1988).

Lithogeochemical data indicate that the biotite monzogranite phase of the Davis Lake Complex is geochemically similar to the 'metal-specialized' granites defined by Tischendorf (1977; see Table 1). Two noticeable geochemical characteristics of the Davis Lake Complex in regard to lithogeochemical exploration is its very low B content and P enrichment, the latter increasing progressively from 'parental' biotite monzogranite to the more altered greisenized phases of the complex (Table 1).

The distribution of tin in bedrock of the Davis Lake Complex and associated Meguma metasediments is shown in figure 4. The enrichment of tin in the Carapace Zone outlined in this figure is evident with the highest concentrations occurring in areas where greisenization is present.

GEOCHRONOLOGY

Geochronological interpretations of the age of emplacement of the Davis Lake Complex, the timing of mineralizing events, and temporal relationships of tectono-thermal and deformational events in this area are still quite controversial; the principal workers in this field being Richardson et al, (1990) and Kontak and Cormier, (1990, in press). Interpretations given by these two sets of workers are presented here with the understanding that scientific rebuttals on this subject await final publication of data.

'Richardson et al, (1990)':

The extremely high Rb and low Sr contents of the Davis lake Complex (DLC) are typical of "specialized" granites (Tischendorf, 1977; see Table 1), and there resultant Rb/Sr ratios are at least three times those found in "barren" granites. All whole-rock data (n = 8) for the DLC yield a scatterchron with a date of 336 $\pm$ 14 Ma (MSWD=24) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.722 $\pm$ 0.002. If data from samples close to the contact are excluded from the calculation, a five-point isochron results that yields a date of 330 Ma $\pm$ 7 Ma (MSWD 2.8) and an initial ratio of 0.727 $\pm$ 0.004.

Metawacke in the DLC area 330 Ma ago had an 87Sr/86Sr of about 0.713. The presence of partially assimilated metasedimentary xenoliths near the granite metawacke contact indicates that, in this area, the DLC magma assimilated Sr from the metawacke that was less radiogenic than that already in the magma. Such assimilation could create isotopic heterogeneity and result in a high MSWD. If closed-system conditions have existed since 330 Ma, samples contaminated with metawacke should fall below the isochron because these samples have a primary less radiogenic composition, as do those samples located close to the upper contact. If these samples are added separately to the isochron, the MSWD increases significantly.

The omission of the data from EK49, a sample of biotite-chlorite-muscovite monzogranite, from the isochron calculations results in only a slight improvement in the MSWD (2.8 to 2.6). This indicates that the late-magmatic fluids responsible for the alteration of biotite to chlorite in the DLC were in isotopic equilibrium with both the crystallized minerals and the crystallizing magma.

Richardson et al., (1987), reporting on a subset of this data, did not reject any data on geological grounds. Accordingly, the results they presented had a high MSWD that was similar to that quoted for n = 8 samples. This high MSWD was attributed to primary isotopic heterogeneity throughout the DLC generated as a result of the inheritance of Sr from biotite granite xenoliths. Although these xenoliths could have contributed inherited Sr to the DLC magma, the isochron relationship indicates that the interior of the Davis Lake pluton was isotopically homogeneous with respect to Sr.

The DLC biotite monzogranite is highly evolved, Rb-rich and Sr-poor compared with other Appalachian granites. The DLC biotite contains four times more Rb and four times less Sr when compared with biotite from other Appalachian granites. In the absence of an aqueous fluid, the high Rb content of undersaturated DLC magma appears to have been locally accommodated in biotite. Given the extreme Rb/Sr ratios (about

DAVIS LAKE PLUTON (Devonian–Carboniferous)

Biotite Monzogranite (aplitic near contact)

Biotite Monzogranite, Biotite-Muscovite Monzogranite, Biotite-Muscovite-Chlorite Monzogranite

Carapace Zone: Muscovite Monzogranite, Greisenized Monzogranite (blue and white/yellow leucogranite), minor pegmatite

Greisen Ore

Albitite (± magnetite)

Kaolinized/Silicified granite and brecciated metasediments

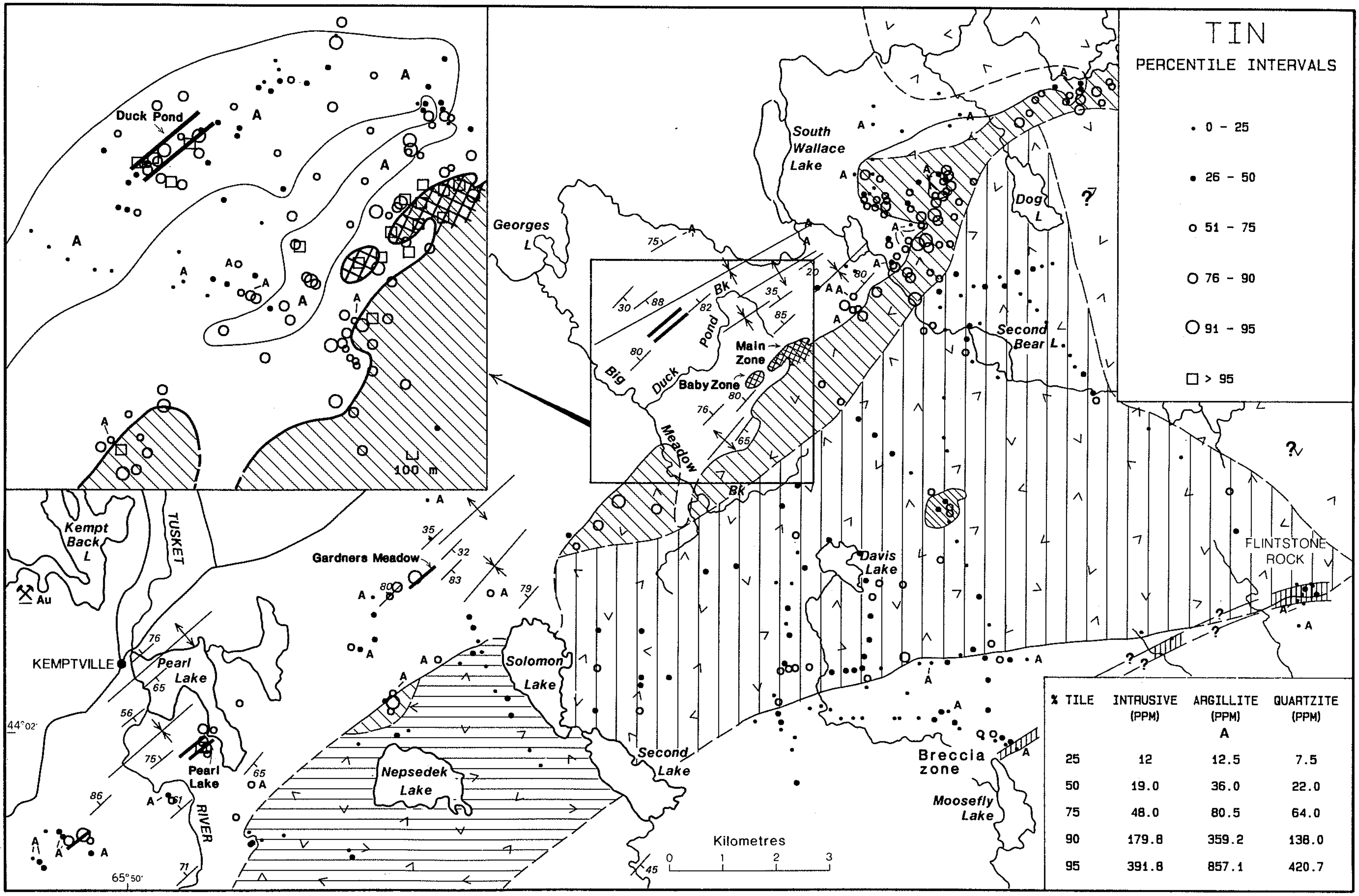
MEGUMA GROUP (Cambro–Ordivician)

Goldenville Formation (meta–wacke, meta–argillite–A)

Cassiterite–sulphide vein, stratiform and breccia zones in Goldenville Formation

Intrusive–metasediment contact (defined, assumed) . .

Legend for Figure 4.



% TILE	INTRUSIVE (PPM)	ARGILLITE (PPM) A	QUARTZITE (PPM)
25	12	12.5	7.5
50	19.0	36.0	22.0
75	48.0	80.5	64.0
90	179.8	359.2	138.0
95	391.8	857.1	420.7

Figure 4. Representative analyses of the main types of greisen mineralization in the East Kemptville Tin Deposit (after Richardson et al., 1990a)

Table 1. Average compositions (arithmetic and geometric means) of the various rock types comprising the Davis Lake Pluton (oxides in wt%; trace elements in ppm except where noted; greisen ore samples represent splits of 100 tonne, 3 M^3 blocks of underground ore through main drift; after Boyle unpublished data).

	BIOTITE MONZOGRANITE (N = 129)		BIOT-MUSC MONZOGRANITE (N = 48)		MUSCOVITE MONZOGRANITE (N = 23)		GREISENIZED MONZOGRANITE (N = 41)		GREISEN ORE (N = 43)		ALBITITE (N = 10)		KAOLINIZED MONZOGRANITE (N = 15)		SPECIALIZED GRANITE (TISCHENDORF, 1977)
	AM (SD)	GM	AM (SD)	GM	AM (SD)	GM	AM (SD)	GM	AM (SD)	GM	AM (SD)	GM	AM(SD)	GM	
SiO2	75.81 (1.62)	75.83	75.88 (1.21)	75.84	76.16 (1.68)	75.86	74.24 (3.49)	74.13	73.00 (.92)	72.44	70.92 (4.11)	70.79	90.81 (9.44)	91.20	73.38 ± 1.39
TiO2	.15 (.06)	.13	.12 (.04)	.09	.11 (.07)	.09	.09 (.13)	0.06	.04 (.01)	.04	.12 (.06)	.09	.10 (.12)	.06	.16 ± 0.10
Al2O3	13.01 (.68)	12.88	13.19 (.60)	13.18	13.07 (.82)	13.18	14.13 (1.14)	14.13	14.55 (.66)	14.45	15.20 (2.57)	15.14	5.08 (4.67)	2.95	13.97 ± 1.07
FeO	1.07 (.38)	1.00	.99 (.34)	.74	.90 (.46)	.74	1.80 (.56)	0.58	2.70 (.78)	2.57	.40 (.28)	.35	.69 (.33)	.62	1.10 ± 0.47
Fe2O3	.99 (.71)	0.69	.88 (.65)	.71	.96 (.54)	.71	1.58 (1.37)	0.66	.29 (1.03)	.02	2.71 (2.42)	2.14	.52 (1.16)	.05	.80 ± 0.47
Fe2O3T	1.66 (.43)	1.62	1.52 (.29)	1.45	1.48 (.28)	1.45	2.08 (1.08)	1.91	3.06 (.87)	2.95	2.81 (2.35)	2.34	1.09 (1.15)	.78	
CaO	.54 (.21)	.51	.46 (.11)	.45	.48 (.11)	.47	.53 (.13)	0.50	.79 (.15)	.78	.43 (.15)	.41	.12 (.19)	.07	.75 ± 0.41
MgO	.24 (.22)	.18	.16 (.10)	.13	.17 (.09)	.13	.22 (.40)	0.15	.07 (.05)	.05	.48 (.34)	.42	.20 (.37)	.06	.47 ± 0.56
Na2O	2.94 (.55)	2.88	3.07 (.39)	3.02	3.03 (.59)	2.95	2.67 (.98)	2.40	2.06 (.73)	1.86	7.70 (2.30)	7.41	.45 (.97)	.43	3.20 ± 0.61
K2O	4.62 (.47)	4.57	4.55 (.33)	4.57	4.11 (.31)	4.07	3.77 (.70)	3.63	3.48 (.44)	3.47	1.28 (1.11)	.98	1.21 (1.60)	.45	4.69 ± 0.68
P2O5	.15 (.07)	.14	.19 (.08)	.17	.21 (.10)	.19	.29 (.13)	0.26	.48 (.04)	.48	.16 (.08)	.15	.04 (.06)	.03	
CO2	.13 (.07)	.12	.13 (.05)	.12	.13 (.06)	.12	.12 (.04)	0.11	.16 (.06)	.15	.16 (.07)	.14	.12 (.04)	.11	
S	100 (364)	52	122 (203)	69	116 (111)	83	1642 (2344)	646	4954 (2931)	4266	605 (1550)	123	86 (36)	78	
F	1719 (765)	1549	1979 (823)	1818	3648 (1381)	3388	7119 (4120)	6026	15758 (3449)	15488	930 (853)	661	118 (128)	74	3700 ± 1500
Cl	261 (117)	240	281 (124)	257	309 (102)	295	334 (132)	309	566 (165)	550	368 (93)	335	138 (101)	98	
Br	2 (2)	2	3 (2)	3	5 (3)	4	5 (4)	4	6 (2)	6	5 (0.0)	5	1 (.9)	1	
B	18 (7)	17	20 (9)	17	17 (5)	17	17 (3)	17	12 (4)	11	21 (8)	19	7 (3)	6	
Li	118 (64)	100	150 (75)	129	281 (144)	240	385 (174)	309	683 (209)	741	51 (44)	37	26 (15)	22	400 ± 200
Rb	439 (107)	427	480 (99)	468	664 (170)	645	787 (241)	741	985 (113)	977	265 (222)	118	99 (113)	47	580 ± 200
Cs	14 (5)	12	16 (5)	15	23 (8)	21	25 (11)	23	32 (7)	32	12 (12)	8	5 (3)	4	
Sr	31 (33)	22	20 (12)	17	28 (25)	20	51 (105)	23	40 (13)	38	57 (77)	30	37 (43)	20	
Ba	190 (135)	162	146 (93)	126	129 (97)	107	139 (196)	98	33 (13)	31	107 (42)	100	254 (286)	135	
Be	7 (3)	6	8 (6)	7	8 (5)	7	5 (4)	4	9 (4)	8	7 (5)	6	.6 (.3)	.6	13 ± 6
Sn	17 (14)	14	27 (18)	23	54 (26)	49	635 (1038)	263	1634 (830)	1479	60 (93)	26	8 (4)	7	40 ± 15
W	10 (11)	6	18 (42)	9	75 (291)	16	27 (25)	16	72 (60)	56	1 (.5)	1	5 (10)	3	7 ± 3
Nb	14 (3)	14	16 (4)	16	23 (7)	22	33 (14)	30	30 (3)	30	16 (6)	15	5 (4)	3	
Ta	3 (1)	2	3 (2)	3	5 (3)	5	7 (4)	6	12 (2)	12	3 (1)	3	2 (.7)	2	
V	10 (6)	7	8 (5)	6	11 (18)	6	4 (4)	2	11 (7)	8	9 (7)	6	11 (15)	7	
Sc	2.4 (1.0)	2.3	2.2 (.7)	2.1	2.3 (.7)	2.2	2.4 (.9)	2.2	1.6 (.2)	1.6	2.0 (.7)	1.9	1.0 (1.2)	5	
Y	30 (19)	21	22 (13)	17	18 (9)	16	14 (11)	10	4 (3)	3	29 (21)	16	4 (7)	2	
La	109 (69)	69	104 (63)	83	55 (83)	17	53 (66)	18	2 (.6)	2	1 (0.0)	1	1 (0.0)	1	
Zr	103 (36)	96	79 (23)	76	81 (54)	68	36 (27)	26	30 (4)	30	155 (76)	132	43 (73)	12	
U	15.9 (6.7)	14.5	16.6 (5.7)	15.5	18.8 (8.0)	16.5	21.9 (6.1)	20.9	25.4 (1.3)	25.1	24.0 (23.2)	17.8	2.9 (3.5)	1.3	
Th	14.5 (5.9)	13.5	11.3 (3.8)	10.5	10.4 (3.4)	10.0	10.1 (9.8)	8.1	4.4 (.6)	4.4	16.2 (5.6)	15.1	2.3 (3.5)	1.3	
Cu	4.8 (6.7)	2.9	13.9 (29.7)	5.1	11.2 (10.0)	8.7	257.3 (460.6)	81.3	770.2 (404.9)	708.0	19.5 (41.1)	4.3	5.9 (5.2)	4.5	
Pb	12 (8)	10	10 (5)	8	8 (5)	6	7 (6)	5	14 (11)	11	5 (4)	3	22 (49)	6	
Zn	75 (471)	33	47 (36)	39	52 (31)	47	558 (995)	214	1674 (1328)	1230	40 (37)	30	27 (43)	13	
Mn	222 (89)	204	212 (59)	204	301 (91)	288	491 (206)	447	626 (136)	617	287 (287)	245	224 (147)	69	
Au (ppb)	1 (.3)	1	1 (0.0)	1	1 (.5)	1	1 (0.0)	1	3 (2)	2	- -	-	- -	-	
Ag	.09 (.25)	.06	.09 (.12)	.07	.12 (.13)	.09	.79 (1.21)	0.34	1.63 (.95)	1.48	.12 (.12)	.09	.07 (.05)	.06	
Cd	.4 (3.2)	.1	.4 (.9)	.2	.2 (.2)	.1	5.6 (9.0)	1.5	16.3 (13.8)	11.2	.1 (.1)	.1	.2 (.3)	.4	
Mo	2.5 (2.5)	1.7	2.3 (3)	1	2 (3)	1	4 (4)	2	6 (4)	4	1 (2)	.93	3 (3)	3	3.5 ± 2
Ni	5 (4)	4	5 (4)	3	3 (3)	2	2 (2)	1	6 (5)	5	3 (2)	2	6 (10)	2	
Co	2 (1)	2	2 (1)	2	1 (.8)	1	1 (1)	1	5 (4)	3	2 (1)	2	3 (4)	2	
Hg (ppb)	47 (198)	23	29 (19)	22	2 (13)	17	14 (7)	12	28 (17)	23	14 (12)	12	17 (9)	15	
As	3.1 (4.5)	2.2	6.9 (18.4)	2.7	7.2 (18.7)	2.7	27.9 (92.7)	6.8	76.4 (61.2)	55.0	5.2 (8.1)	3.0	1.5 (1.0)	1.1	
Sb	.4 (.3)	.3	.3 (.1)	.3	.3 (0.0)	.3	.4 (.3)	0.3	1.1 (.8)	.8	.5 (.3)	.4	.3 (0.0)	.3	
Bi	3 (3)	3	4 (5)	3	4 (2)	3	18 (33)	8	31 (20)	26	3 (0.0)	3	5 (4)	4	
Se	.5 (.4)	.4	.5 (.7)	.4	.3 (.1)	.3	.4 (2)	0.4	.4 (.2)	.4	.3 (.1)	.3	.3 (0.0)	.3	
Tl	3 (.4)	3	3 (.5)	3	3.6 (1)	3.5	4 (1.6)	4	5.5 (1.0)	5	3.0 (0.0)	3	3 (0.0)	3	

3000) of the DLC biotite, significant amounts of ^{87}Sr would be evolved by the radioactive decay of ^{87}Rb. This ^{87}Sr could be easily lost during even minor tectonothermal disturbances. Rb-Sr isotopic data for 4 biotite samples do not define an isochron but individual biotite - whole-rock isochrons yield dates between 260-235 Ma which are younger than the K-Ar (Harlow 1981) and $^{40}Ar/^{39}Ar$ determinations from the DLC area (Zentilli and Reynolds, 1985). However, data for whole-rock samples which are part of the DLC isochron, imply that ^{87}Sr was not redistributed beyond the whole-rock scale (1-3 dm3).

Geological constraints do not bracket the crystallization age of the DLC tightly. The complex intrudes the Cambro-Ordovician Meguma Group. Geophysical evidence indicates that the DLC is distinct from, and appears to intrude, the Devonian SMB. Textural observations indicate that the DLC intruded pre- to syn-kinematically in a location now adjacent to the Tobiatic Shear Zone thought by Giles (1985) to have been active from 296-330 Ma. $^{40}Ar/^{39}Ar$ studies on muscovite in the East Kemptville deposit yield dates of 295 + 5 Ma (Zentilli and Reynolds, 1985).

The Rb-Sr isotopic system appears to have remained closed at the whole rock scale in the DLC, but not at the mineral scale. The isochron whole-rock date of 330 + 7 Ma, determined on DLC biotite monzogranite, is much greater than the dates determined from biotite - whole-rock sample pairs (260-236 Ma). The 330 Ma - early Carboniferous (Namurian-Visean) - whole-rock date could record either: (i) the time of emplacement and crystallization of the DLC, (ii) the time of complete resetting of the Rb-Sr system, or, less probably, (iii) a time of partial resetting.

It seems most probable that the date indicated by the Rb/Sr isochron preserved in the DLC biotite monzogranite reflects a primary crystallization age. This interpretation does not conflict with the field relationships, textural observations or other chronologic determinations. The DLC appears to have been intruded immediately prior to initiation of movement along the Tobiatic Shear Zone. The $^{40}Ar/^{39}Ar$ date of 295 + 5 Ma on DLC muscovite coincides with Giles's (1985) timing for the end of movement in this shear zone.

If the DLC were older than 330 Ma (perhaps the same age as the SMB), then the whole-rock isochron reflects the effects of resetting. In this case, the high initial $^{87}Sr/^{86}Sr$ ratio of the DLC would result from redistribution of radiogenic Sr generated between the times of intrusion and resetting. This proposal would require redistribution of radiogenic Sr over virtually the entire extent of the DLC. There is no petrographic or geochemical evidence for the introduction of post-crystallization fluids into the DLC. Thus any chronologic resetting must have occurred by dry thermal or mechanical means and must have been restricted in scale. Given the widespread geographic distribution of the large whole-rock samples used in this study, it is difficult to support such a model.

The open-system behaviour of DLC biotite is attributed to tectonothermal disturbances in the Meguma Terrane documented by $^{40}Ar/^{39}Ar$ geochronology.

Whole-rock Rb-Sr isotopic data from eight geographically widespread samples of greisen yield 317 $\pm$ 32 Ma (MSWD = 25) and an initial $^{87}Sr/^{86}Sr$ of 0.733 $\pm$ 0.010. When the 5 samples with ^{87}Rb ^{86}Sr less than 100 are used, an isochron results that yields 337 $\pm$ 5 Ma (MSWD = 2.9) and an initial $^{87}Sr/^{86}Sr$ of 0.729 $\pm$ 0.001 (Richardson et al. 1988a). These ages and initial ratios are indistinguishable from those of the DLC biotite monzogranite.

The high MSWD reflects either loss of radiogenic Sr (Sr*) due to tectonothermal events in the Meguma Terrane, (Richardson et al. 1989) or the interaction of greisen fluid with Sr*-rich minerals in the metawacke (see Richardson et al. 1990a).

'Kontak and Cormier, (1990,in press)':

The results of a detailed Rb-Sr and $^{40}Ar/^{39}Ar$ geochronological study of the East Kemptville leucogranite, host rock to the East Kemptville tin deposit, have revealed a complex history involving the superimposition of several tectono-thermal events after emplacement and crystallization. Although the present study has not unequivocally determined the age of intrusion, it has provided important insight into the conflicting results of previous Rb-Sr and $^{40}Ar/^{39}Ar$ studies. The results of our study, when combined with previously published data, indicate the following chronology of magmatic-tectonic-thermal events for the East Kemptville leuco-granite and immediate area:

(1) Intrusion and mineralization at ca. 370 Ma based on results of Pb isotope analyses of A.K. Chatterjee (pers. commun.). This age is the same as that obtained for other phases of the South Mountain Batholith by a variety of methods (U/Pb, Rb/Sr, $^{40}Ar/^{39}Ar$) and indicates that the East Kemptville leucogranite represents a time equivalent.

(2) Closure of the whole rock Rb-Sr and K-Ar muscovite systems at ca. 353 Ma based on an eleven point Rb-Sr whole rock isochron age of 353 6 Ma and the high temperature plateau segment of a $^{40}Ar/^{39}Ar$ age spectrum for muscovite (determined by Zentilli and Reynolds, 1985). This age is considered to reflect a pervasive tectono-thermal event, but it is not possible to say if it represents one of several episodic events or the termination of a single protracted episode of deformation.

(3) A seven point Rb-Sr muscovite isochron age of 334 + 2 Ma reflects another overprinting event which was of sufficient magnitude to cause open system behaviour and resetting of muscovites but not the whole rocks. The preservation of an older Rb-Sr whole rock isochron age indicates that isotopic redistribution was restricted to a minimum volume represented by the whole rocks in this study. This ca. 330 Ma event is also corroborated by two additional lines of evidence, namely a 338 Ma plateau age for the high temperature segment of a $^{40}Ar/^{39}Ar$ age spectrum obtained on muscovite and a previously published (Richardson et al., 1988) whole rock Rb-Sr isochron age of 337 $\pm$ 5 Ma for greisen samples. The latter age is considered to reflect the dominance of muscovite in the analyzed samples, thus duplicating the data obtained on Rb-Sr analyses of muscovite separates in this study.

(4) Evidence of a ca. 300 Ma thermal event is suggested by two previously published $^{40}Ar/^{39}Ar$ muscovite age spectra (Zentilli and Reynolds, 1985). The only additional evidence found for this event during the present investigation was in the low temperature part of the muscovite $^{40}Ar/^{39}Ar$ age spectrum which indicated an overprinting event of ca. 280-300 Ma age.

(5) Isotopic redistribution in feldspar phases at ca. 260 Ma is recorded by seven Rb-Sr WRMMS-QP-KF isochrons. It is not possible to say if this event represents a later superimposed event of the waning stages of an earlier event.

Evidence for the regional extent of the aforementioned events is found in previously published geochronological data for the southern Meguma Terrane. In particular we cite (1) the 330 $\pm$ 7 Ma Rb-Sr whole rock isochron age and ca. 260 Ma Rb-Sr whole rock-biotite ages for the Davis Lake Pluton (Richardson et al., 1987, 1989), (2) the ca. 315 Ma whole rock Rb-Sr and U-Pb zircon ages for the Wedgeport granite (Cormier et al., 1988), (3) the numerous $^{40}Ar/^{39}Ar$ ages obtained throughout the southern Meguma Terrane, and (4) the young Rb-Sr whole rock ages obtained at Long Lake (332 $\pm$ 10 Ma) and Westfield (270 $\pm$ 14 Ma) by O'Reilly et al. (1985).

The above documentation of numerous tectono-thermal events, the presence of deformational fabrics within the East Kemptville area and the regional extent of the EKSZ and other shear zones (RLSZ, TFZ, EKEDSZ; Fig. 2 and 3) indicate that protracted tectonism characterized the Late Devonian-Carboniferous period in the southern Meguma Terrane. We note, however, that it is difficult to say whether these features record episodic tectonism or localized closure of the isotopic systems to a protracted event.

DAVIS LAKE COMPLEX (DLC)

Geological Setting

The Davis Lake complex (DLC), a suite of biotite monzogranite, leuco-monzogranite and altered granitic rocks in southwestern Nova Scotia, lies within the "suspect" Meguma terrane (Fig. 1) The DLC is poorly exposed (6 outcrops/150 km^2 in the southern part), but drift prospecting, systematic mapping of glacial overburden, and drilling in 1978-1982 revealed that the complex is much smaller than previously suggested by Taylor (1967) and Keppie (1979).

Originally, the DLC was considered part of the adjacent South Mountain batholith (SMB) because of its location, granitic lithology, metal- and volatile-rich nature (McAuslan et al. 1980; Richardson et al. 1982; Chatterjee and Muecke 1982; Chatterjee et al. 1983, Chatterjee and Strong 1984; Clarke and Muecke 1985; MacDonald and Clarke 1985; Ford and O'Reilly 1985) and major and trace-element geochemistry (Richardson 1983a; 1985). Chatterjee et al. (1983) considered the complex to be a Sn-mineralized "para-intrusive" (Chatterjee and Muecke 1982) end-member of a cogenetic suite of SMB granitic rocks. However, regional radiometric data (Ford and O'Reilly 1985) and a compilation of gravity and magnetic data (Egner 1987) suggest that the DLC may be a separate pluton about 45 km long (NE-SW) and 10 km wide (Fig. 1). Petrography and whole-rock geochemistry indicate the DLC is probably a single-phase intrusion that was extensively modified by late-stage magmatic and hydrothermal processes (Richardson 1983a). K-Ar and $^{40}Ar/^{39}Ar$ age determinations on muscovites (Harlow 1981; Zentilli and Reynolds 1985) indicate that the DLC may be as much as 65 Ma younger than the SMB. Kontak (see Geochronology section) has suggested, however, that the DLC is a time equivalent of the SMB.

The contact of the DLC with metawacke is sharp and is characterized by a chilled margin that is 2-5 cm wide and contains large quartz and plagioclase phenocrysts. Large (10-300 m) roof pendants and small, partially assimilated, xenoliths occur near the southwestern contact. Small granitic dikes are present, but pressure-release unidirectional solidification textures do not occur. The emplacement of the DLC appears to have been a relatively passive event, typical of H_2O-poor, F-rich granitic magmas (Pollard et al. 1987).

The DLC contains several textural and mineralogical subtypes of biotite-bearing monzogranite. Bluish-grey, porphyritic Nepsedek Lake biotite monzogranite is composed of white perthitic microcline, quartz and subhedral plagioclase (An0-15) megacrysts that are set in a massive, medium-grained groundmass of potassic feldspar, plagioclase, quartz and biotite. Primary muscovite and topaz are rare, sericitized perthitic microcline crystals contain fluorite (Richardson 1988a). Fluorine and Cl contents of biotites range from 0.49% to 1.06% and 0.11% to 0.26% respectively and have an average F/Cl ratio of 4.

Aplitic biotite monzogranite is a fine-grained subtype of biotite monzogranite. On fresh surfaces, aplitic biotite monzogranite is light-brown in color and consists of a fine- grained, unfoliated groundmass and a phenocryst assemblage that is similar to that of the biotite monzogranite. Biotite in this rock-type is unaltered and is associated with ilmenite, apatite, zircon, and possibly other radioactive minerals. Biotite-chlorite- muscovite (bt-cl-mu) monzogranite, a mappable unit defined by the replacement of biotite by chlorite and/or muscovite is foliated locally. Perthitic microcline takes on a pink or greyish-green color where it has been subjected to hematitic or sericitic alteration. Many microcline crystals in this rock-type have albite rims.

Green greisen zones (2 to 5 cm wide) occur along fractures in aplitic biotite monzogranite and in bt-cl-mu monzogranite. Biotite and feldspar are completely replaced by green mica and minor fluorite; quartz is partially overgrown by mica. Sulphide minerals, cassiterite and topaz do not occur in this greisen type. An outer hematite-rich zone that is 5 to 10 cm wide envelopes an inner zone in which 60% to 75% of the feldspars are replaced. Grey greisen zones occur beneath the southwestern granite-metawacke contact. In contrast to the green greisen zones, these 3 to 10 cm bands envelope fractures or quartz veins that contain fluorite, cassiterite and/or base-metal sulphide minerals. Feldspar that occurred marginal to grey greisen zones is replaced by grey mica, but there is no outer hematite envelope. Biotite and quartz are present in grey greisen.

Several types of xenoliths and unusual phenocrysts occur in the DLC. Biotite granite xenoliths (1-30 cm) can have successive rims of biotite and sericitized plagioclase or show marginal assimilation. Fractured, bluish-colored quartz megacrysts contain biotite, unaltered microcline, muscovite, plagioclase and zircon. Glomeroporphyritic biotite contains cheralite, fluorite, fluoroapatite, ilmenite, muscovite, zircon and a rare-earth element (REE) fluoride (?parisite). Near the contact, partially assimilated metawacke xenoliths are present.

Locally within the DLC, perthitic microcline is aligned, presumably preserving magmatic flow directions. Deformation textures such as undulose extinction, subgrain development and bent twin lamellae are ubiquitous in the DLC. Incipient gneissosity, mosaic quartz, kink-bands and mylonite occur at the southeast contact, near the postulated trace of the Tobiatic Shear Zone. Kontak (1987) documented mylonite and C-S fabrics in the East Kemptville deposit and the granite-metawacke contact is faulted at this location (Richardson et al. 1988).

Major and Trace-Element Geochemistry

Outside of the Carapace Zone on figure 4, major-element trends in the DLC are limited in extent, but generally parallel those expected with the fractional crystallization of granitic minerals (e.g. trends also characteristic of the SMB). However, in the DLC, P_2O_5 decreases with increasing SiO_2. DLC rocks are more siliceous and oxidized, but contain less alumina. Chemical variation within the mineralogical and textural varieties of DLC biotite monzogranite is restricted. Aplitic biotite monzogranite is generally indistinguishable from coarser-grained biotite monzogranite and all varieties of DLC monzogranite overlap on an Ab-An-Or plot. On a Q-Ab-Or plot, most DLC data cluster near the ternary minima (Tuttle and Bowen 1958; Manning 1981) for water-saturated granite at 0.5 and 1 kb, indicating crystallization occurred at a shallow depth. Normative trends associated with common types of granitic alteration (c.f. Stemprok 1976) are not defined, implying many of these samples of the DLC have retained their primary geochemical signature.

Thornton-Tuttle differentiation indices (D.I.) are high for the DLC samples (92 to 94). The low TiO_2 (0.29 to 0.08 wt.%) and Zr (40 to 155 ppm) abundances indicate that these rocks are highly evolved. The biotite-chlorite-muscovite monzogranite appears to be the most differentiated and the aplite biotite monzogranite (marginal to the contact, Fig. 1), is the least differentiated of the DLC biotite-bearing monzogranites. Although Zr is thought to be mobile in F-rich environments (Bandurkin 1961; Alderton et al. 1980; Muecke and Clarke 1981), there is little scatter in the distribution of the data points, again implying that the DLC biotite monzogranites were not affected by F-rich fluids. Although the DLC array has a slightly different slope from that defined by the SMB data, the suites are not easily distinguishable.

The DLC is peraluminous and contains normative corundum. Most biotite monzogranite samples from the southern part of the complex have $c<1.5$, values which are lower than those of the SMB. Aplitic biotite monzogranite and bt-cl-mu monzogranite from the DLC and high-SiO_2 rocks (i.e. biotite monzogranite) of the SMB have $1.5<c<2.5$. Normative corundum increases with the transfer of alkali elements into a coexisting aqueous fluid (Cawthorn et al. 1976; Clarke 1981). Thus, the distribution of high c values in the DLC implies that fluid saturation was geographically restricted to areas near the southwestern granite-metawacke contact and again that portions of the DLC did not have significant interaction with an aqueous fluid during initial stages of emplacement.

Chondrite-normalized, REE patterns for the DLC biotite monzogranite and aplitic biotite monzogranite are relatively flat with Ce/Yb = 10 (Richardson et al. 1983b). The LREE contents of SMB rocks and DLC biotite monzogranites overlap, but the HREE (heavy rare earth elements) contents of the DLC are significantly greater. The constant, moderate negative Eu anomaly present in the DLC data, the limited variations in Ca, Na and K, and Rb/Sr versus Zr+Ce+Yb indicate that feldspar fractionation was not as significant a process during the crystallization of the DLC as it was in the SMB. The limited abundance of Zr in the DLC indicates that although zircon fractionation occurred, this also was not an important process during the differentiation. The enrichment in HREE in the DLC rocks with respect to the SMB biotite granodiorite was therefore not likely related to zircon crystallization.

Elemental abundances of the DLC biotite-bearing monzogranites are listed in Table 1. All units contain abundant Sn and F. In contrast to the SMB, the DLC does not contain tourmaline and is unusually low in B (<25 ppm B). Sn, W, Mo, Li, F and Rb contents are enriched 2 to 13 times the contents of these elements in "lithophile-barren" granites and thus the DLC is "metal- specialized" (Tischendorf 1977), as is much of the SMB. The DLC monzogranites are considered potentially metalliferous using the numerous elemental ratios derived from Sn, W, Mo, Li, F, Rb and high field-strength elements (Flinter et al. 1972; Tauson and Koslov 1973; Smith and Turek 1976).

Both the SMB and DLC are enriched in Sn with respect to most granites. The average DLC Sn abundance is similar to that of the bulk of the SMB suite. Overall, compared to the SMB suite, the DLC biotite monzogranites contain less Ti, again suggesting the DLC rocks are more differentiated. The low-Ti members of the SMB suite are enriched in Sn compared to the majority of this suite. The Sn enrichment in the SMB was attributed by Clarke and Muecke (1985) to the partitioning of Sn into late-stage aqueous fluids generated by the more evolved magma.

Elemental distributions within the DLC (Fig. 4 for Sn; Boyle 1988; Boyle unpublished data) display distinct enrichment-depletion patterns which define a Carapace Zone shown in figure 4 that is dominated by muscovite monzogranite and variably greisenized monzogranite (see Table 1). Richardson (1988a) believes that these elemental enhancements and depletions and their geographical restriction to the Meguma-granite contact area are due to processes that occurred prior to crystallization, implying that the DLC magma was zoned prior to crystallization, a process probably aided by the enhanced diffusion and depolymerization resulting from high F and P contents. The depression of the solidus and liquidus temperatures due to the F content would allow these processes to continue longer than usual. Richardson (op cit) cites such mechanisms as gravitational diffusion, convective diffusion, convective fractionation and side-wall crystallization (see references op cit) as possible processes to account for the observed zoning. Boyle (1990) on the other hand believes that the Davis Lake Pluton and its associated Carapace Zone is characterized by progressive hydrothermal alteration of a parental biotite monzogranite to form biot-musc monzogranite, musc-monzogranite, greisenized monozogranite and greisen stockwork (Table 1). This sequence displays, on a mass balance adjusted basis (Gresen's Formula), marked progressive increases in Sn, Cu, Zn, Cd, As, Sb, Bi, Ag, Li, Rb, Cs, F, Cl, P, Fe, Mn, Mo, W, Nb, Ta, Co, Ni, Be, Tl, U, Sr, Ca, Al and CO_2; no discernable changes in Si, V, Pb, Hg, and Se; and marked progressive decreases in K, Na, Mg, Ti, Th, Sc, B, Zr, Ba, Y, and La. Further evidence to support formation of this Carapace Zone as the result of hydrothermal alteration is A). the absence of primary muscovite in this zone, B). the progressive alteration of biotite (and K feldspar) to muscovite and pyrite (requiring introduction of S) and concomitant movement of Sn from the biotite phase to formation of cassiterite (with further addition of Sn), C). continuation of hydrothermal activity through this zone into the Meguma metasediments to form vein and stratiform deposits (Fig. 4) as well as highly anomalous concentrations of elements characterized by the trend described above in both the fractured and massive portions of the metasediments at the Meguma-granite contact, D) the presence of a sharp granite-sediment chill margin (1-3cm) along the entire length of this zone, including the East Kemptville tin deposit, and the near absence of granitic dyke rocks in the Meguma indicating rapid solidification of the magma at the contact under high confining stresses resulting in formation of a highly interconnected orthogonal fracture system capable of significant fluid transport and resultant formation of greisen veins, stockworks and alteration of massive granite E). geochronological evidence indicating at least three tectono-thermal (hydrothermal?) events in this area (see Geochronology Section), F). the limited fractionation trends shown for the DLC biotite- monzogranite phases of the DLC indicating that fractionation processes were not dominant at the present erosional level. G) the association of this Carapace Zone, tin mineralization within it, and Meguma-hosted hydrothermal tin deposits with a proposed major tectonic structure (East Kemptville-East Dalhousie Shear Zone of Kontak, 1987a, 1987b) which appears to be dominated by brittle deformation and formation of dilation zones capable of acting both as fluid conduits and sites of mineral deposition, H). the concentration of elements in the Meguma-hosted Duck Pond tin deposit (Sn, F, Cu, Zn, Mn, Ag, Cd) which are also enriched in the Carapace Zone and East Kemptville deposit; and enriched of certain elements in the Carapace Zone but not Meguma mineralization (Li, Rb, U, P) and depletion of other elements from both types of mineralization (Na, Cl, Zr, REE) can be more easily explained by hydrothermal processes than by magmatic zoning.

Albitites

Albitites, although volumetrically minor within the Davis Lake Pluton, form a distinctive rock type which differs considerably in its chemistry and petrology from albitized granites formed during mineralizing processes. Typically this rock is coarse grained and composed primarily of albite, quartz and black biotite with magnetite as the main opague mineral. It may also contain muscovite, phenakite, cassiterite, apatite, fluorite and cryolite interstitial to the main albite-quartz matrix (Chatterjee, 1980; Chatterjee and Strong, 1984). Occasionally this rock unit is brecciated and sheared (Chatterjee, 1980) and invariably occurs within zones of intense jointing and shearing in the Davis Lake Complex. A small mappable unit of albitite occurs in the central portion of the complex (Fig. 4). Compared to the biotite-monzogranite unit, albitites are more aluminous and sodic, depleted in K, enriched in Sn, Cl and Zr and contain markedly lower levels of the 'specialized' elements F, Li, Rb, Cs, and W (see Table 1). Temporal placement of the albitites within the intrusion-mineralizing process has not as yet been determined.

Kaolinized Monzogranite

Within the intrusive portion of the Rushmere Lake shear zone (Fig. 3 and 4) the biotite-monzogranite is intensely altered to a highly silicified-kaolinized rock. The rock varies from massive vesicular quartz with crystalline quartz and kaolinite coating and filling the vesicles to silicified, kaolinized granite breccia. Details of the petrology of both the metasedimentary and granitic rock types found in the Rushmere deformation zone can be found in Smith (1985). The chemistry of the kaolinized monzogranite from this area is given in Table 1. None of the lithophile or chalcophile elements, characteristic of the greisenization processes in the northern contact zone, are enriched in these rocks.

EAST KEMPTVILLE TIN DEPOSIT

Geology and Mineralization

The East Kemptville deposit is located under the contact of the Meguma Group metawacke with the East Kemptville (EK) leuco-monzogranite (Fig.4 and 5). The presence of two large roof pendants indicate that the present erosional surface is close to the granite-metawacke contact throughout most of the deposit. Beneath these pendants, the contact is sharp and undulates (Fig. 6). The pendants are erosional remnants of a flat 500 m-wide "shoulder" on the steeply dipping granite-sediment contact, one of the two flat-lying areas oriented NE-SW in the Davis Lake area.

The deposit consists of the Main and Baby Zones (Fig. 4). In the Main Zone, the most extensive and highest-grade (>0.05% Sn) ore occurs near the erosional surface or beneath roof pendants (Fig. 6). The abundance of greisen and the intensity of alteration decreases to the east and with depth. A grade-cutoff of 0.05% Sn delineates the eastern margin of the deposit. The western margin of the deposit is defined by the main steeply northwest-dipping contact (075°/70°W) with the Meguma sediments. Northern and southern boundaries are placed where the meta-sedimentary roof pendant merge with the main contact. The lower margin, also defined by the 0.05% Sn grade-cutoff, occurs 80 to 100 m below surface. At this depth, greisen is scarce and rarely contains cassiterite. The Baby Zone is an elliptical pipe-like apophysis of quartz-topaz rock that is enclosed at surface by steeply-dipping metawacke, but is contiguous with the larger Main Zone at depth.

At the ore-grade cutoff, blue leucomonzogranite grades upward into finer grained EK leucomonzogranite that normally does not contain xenoliths (Fig.6). Near the ore zones, this rock type is yellowish-green. Kontak (1987) reported mylonitization and the development of "C-S fabrics" in the EK leucomonzogranite and greisen. Quartz is fractured, shows

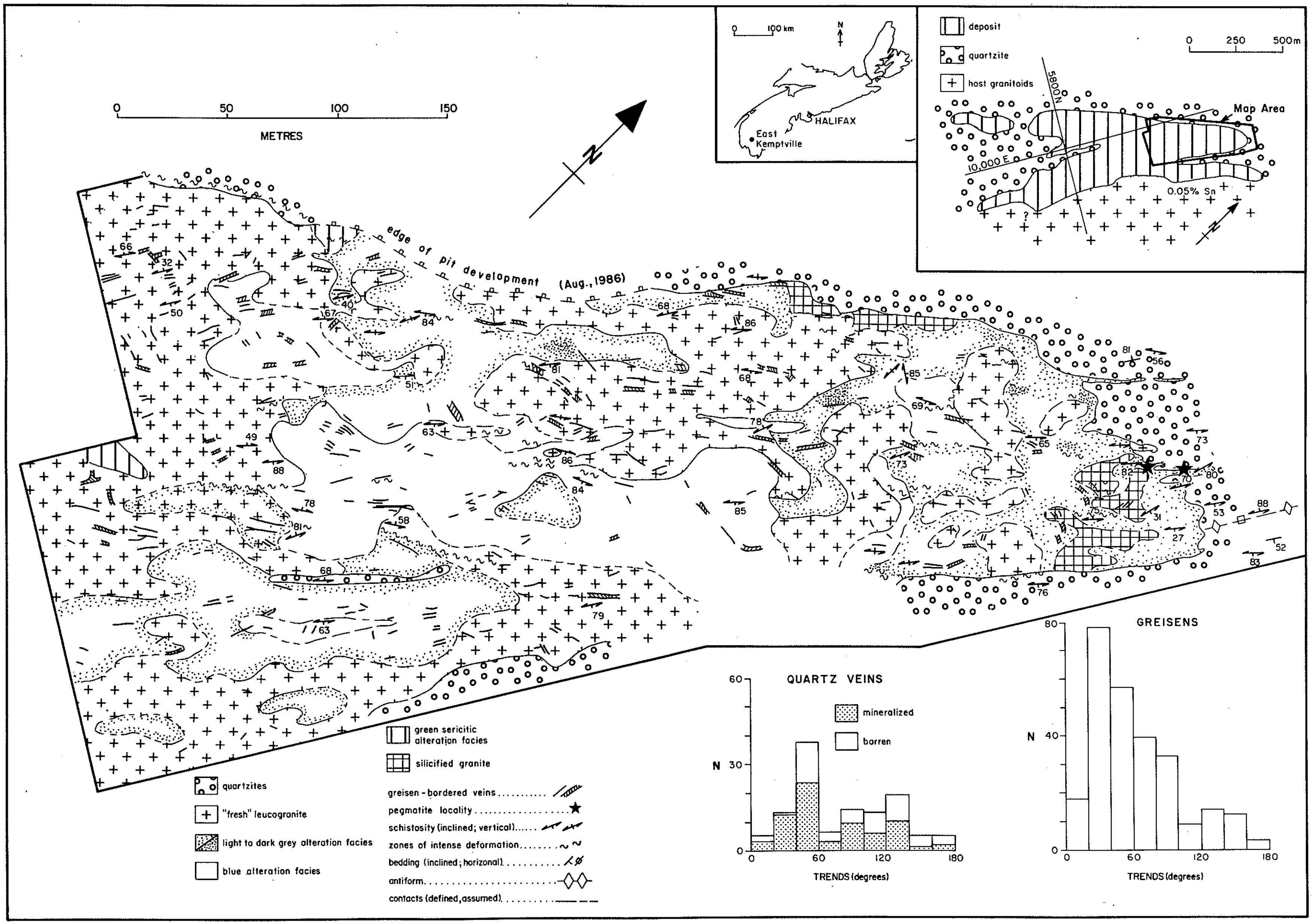


Figure 5. Surface (106 m level) geological map of the northeastern part of the East Kemptville open pit mine. The inset diagram (after Richardson et al., 1982) shows the outline of the deposit as delineated by the 0.05% Sn grade contour and location of the map area. Note that the fabrics shown i the granitoids represent the shear fabric (C plane) only (after Kontak et al., 1986).

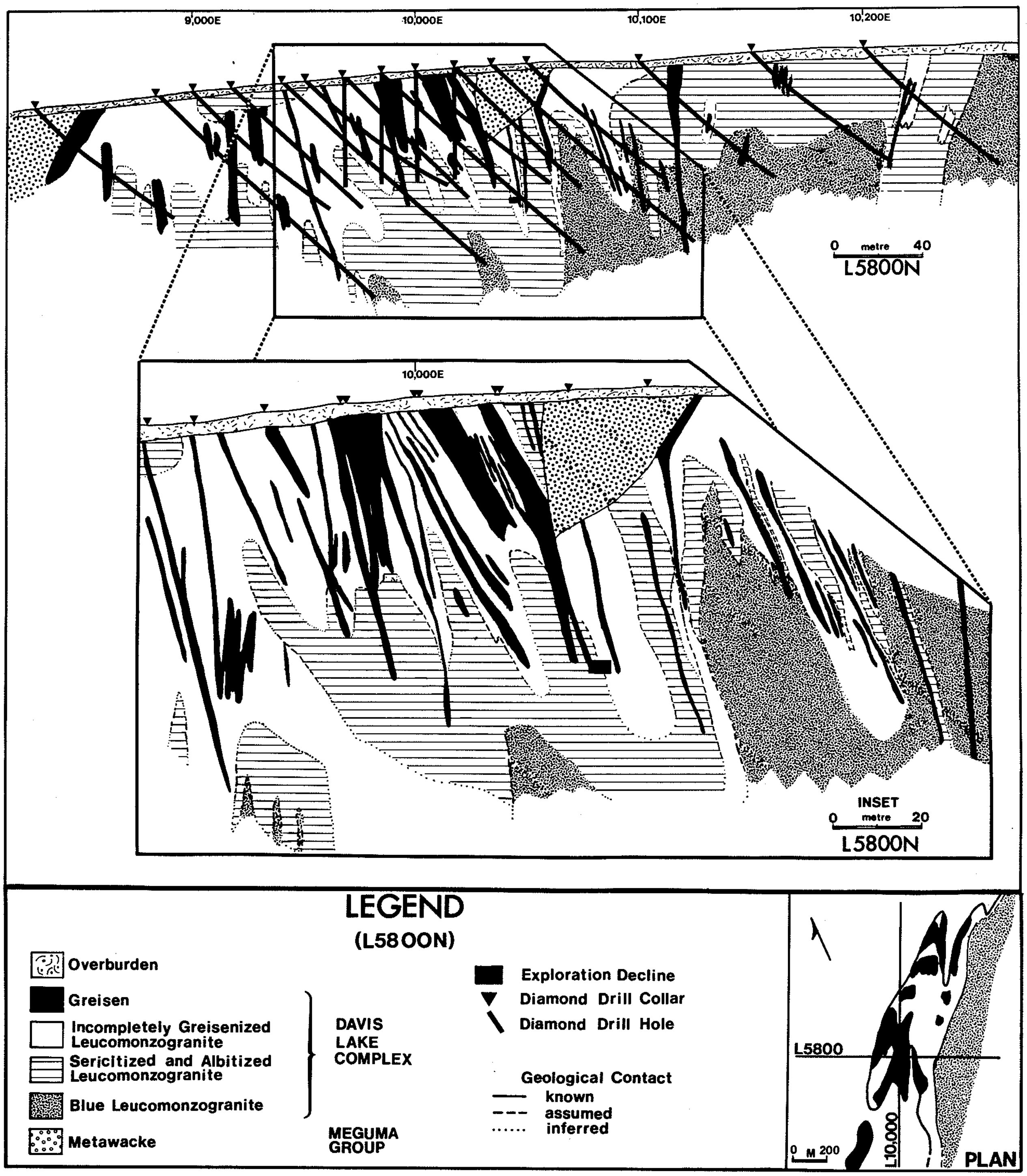


Figure 6. Geological cross section (line 5800N) of East Kemptville Tin deposit (after Richardson, 1988).

granulated grain boundaries, subgrains or undulose extinction as a result. Albite twins are bent or broken. Metawacke, leucomonzogranite and massive greisen near the main granite-metawacke contact (045/75°) are cleaved parallel to the Contact Fault, a 15 cm gouge zone oriented at 046°/84°W that crosscuts the granite metawacke contact (Richardson et al, 1988a). Shearing in the metawacke and cleavage in the granite parallel the contact and one set of greisen bordered zones.

Kontak (1990b) has given the following description to the EK leucomonzogranite: "Petrographically the EKL is a medium grained, equigranular leucomonzogranite characterized by the presence of magmatic topaz and muscovite; rare, small (<.1mm) biotite grains occur. The absence of fluid saturation textures (e.g. miarolitic cavities, UST's, pegmatites) indicate emplacement of the EKL as an undersaturated melt. Although petrographic observations indicate subsolidus modification of primary magmatic mineralogy and textures occurred, the EKL is considered to reflect a magmatic, rather than metasomatic phenomenon. Chemically the EKL is exceptionally uniform and is characterized by elevated values for SiO_2, Al_2O_3(A/CNK=1.13-1.84), total Fe, P_2O_5, Rb, Cs, Ga, Li, F, Sn, Nb, Cu, Zn, W, U, Ta and depletion in TiO2, CaO, MnO, MgO, Ba, B, Th, V, Co, Sc, Hf, Zr, Ni, Mo, Sb, Y. Compared to evolved units of the SMB, the EKL is enriched in Ta, Sr, Rb, F, and Li, while REE abundances and chondritic patterns are markedly different. Geochemically the EKL compares favourably to other F-rich felsic suites, particularly topaz granites.

The EKL is considered to represent the highly fractionated product of a strongly peraluminous melt, itself derived via melting of fluorine-rich biotite from a depleted granulite source region, perhaps the residue from which part of the SMB was previously extracted; it is not considered to represent a fractionate of the SMB (or Davis Lake Complex). The location of the EKL within a major northeast-trending shear zone may indicate that there was a strong structural control in both the generation and emplacement of the EKL".

The overlying metawacke rarely contains granitic dikes or veins that are enveloped by dark alteration zones and contain cassiterite, pyrite, arsenopyrite or muscovite. Where present, both dikes and veins extend only several meters into the metawacke. Skarn is not present. Pressure-quench textures such as those documented by Shannon et al (1982) are not found. Thin fine-grained felsite dikes in the granite are cross-cut by greisen zones (Fig. 7).

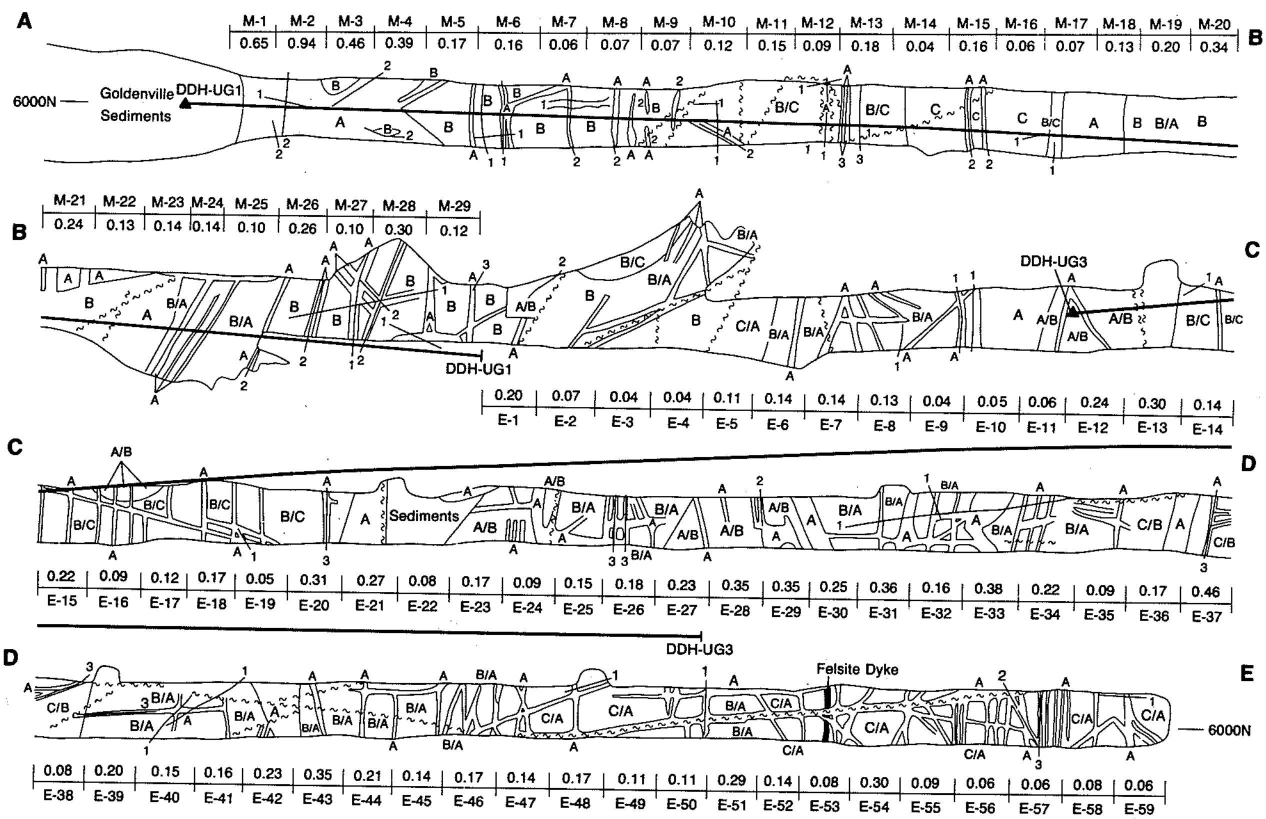
Cassiterite-bearing rocks

Mineral paragenesis in the East Kemptville deposit is summarized in Table 2.

The term "greisen" traditionally refers to altered granitic rocks that consist of F- and B-bearing minerals, quartz and mica and typically cassiterite, pyrite, arsenopyrite, wolframite and sphalerite. Greisen can occur as envelopes bordering veins or as massive greisen zones. The term "greisen" is used here to refer to granitoid rocks in which >90% of the feldspar is replaced by quartz, mica, topaz, cassiterite and other metallic minerals. "Incomplete greisen" is used when 60-90% of the feldspar is replaced (Richardson, 1985; 1988a). The descriptive term "quartz-topaz rock" (Richardson op cit) is applied to a rock that is composed of greisen-style minerals, but shows textural evidence of being a direct precipitate from a F-rich fluid.

At East Kemptville, greisen is most heavily concentrated where the granite-metawacke contact is flat-lying. The highest-grade ore zone in the deposit occurs under the southern roof pendant and is now mined out. Throughout the deposit, small greisen zones occur at the contact. Most of the cassiterite occurs in either symmetrical greisen envelopes around fractures and veins or in larger zones of massive greisen (Fig. 6 & 7). However, cassiterite also occurs in pegmatitic segregations that may or may not be surrounded by small greisen halos. Topaz is always associated with cassiterite, but not the reverse. Boron-bearing minerals such as tourmaline are not observed.

Three types of pegmatitic segregations occur in EK leucomonzogranite and are classified by mineralogy and the presence or absence of alteration envelopes (Richardson et al, 1988a). These features, which occur



LEGEND

A	GREISENIZED LEUCOGRANITE: envelopes with or without central veining (see below for types); 5-30% sulphides comprising pyrite, pyrrhotite, chalcopyrite and sphalerite; disseminated cassiterite
B	SERICITIZED LEUCOGRANITE: containing less than 1% greisen alteration bands (1-2 cm). Minor disseminated pyrite
C	LEUCOGRANITE: weakly sericitized
A/B, B/A, B/C, C/B, C/A	MIXED LITHOLOGIES: e.g. A/B - more A than B

VEIN TYPES

1. White Quartz Veins ± Sulphides, Carbonates, ± Phosphates
2. Grey Quartz-Topaz-Cassiterite
3. Grey Quartz Vein ± Pyrite ± Chalcopyrite ± Sphalerite and ± Cassiterite
4. Sulphide Veins ± Pyrite ± Chalcopyrite ± Sphalerite ± Cassiterite ± Arsenopyrite ± Wolframite (not in main drift)

STRUCTURAL FEATURES

Fracture with open cavities .... ~~~~

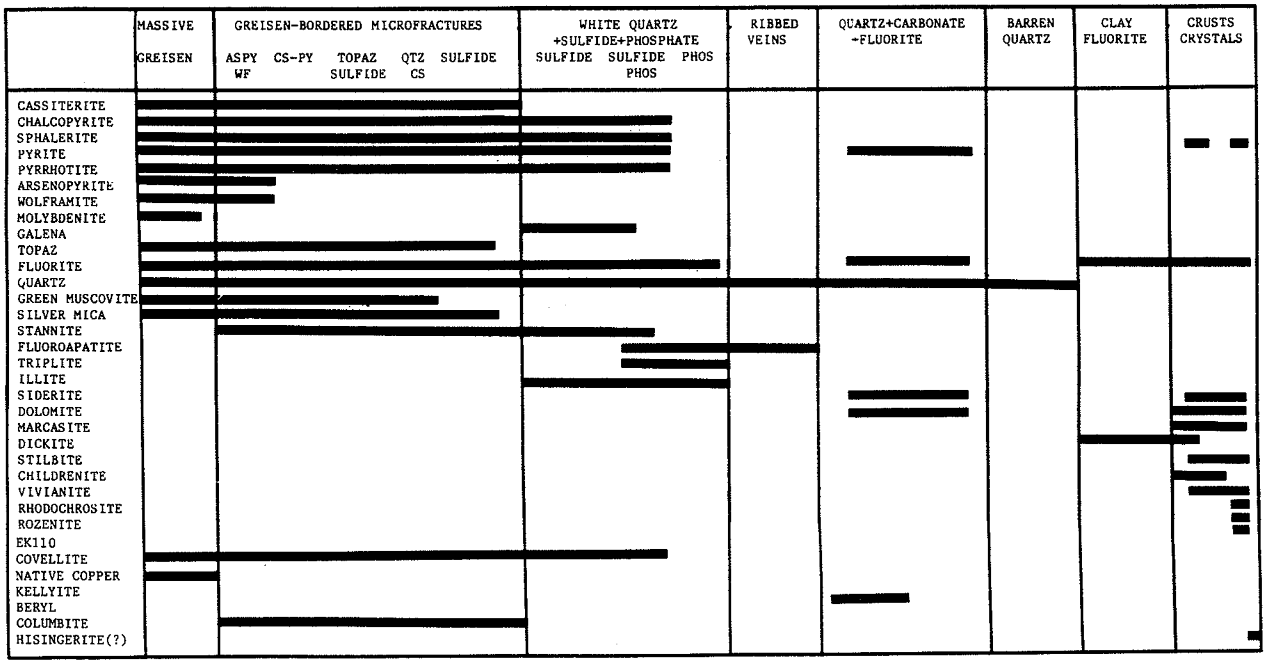
Contact, sharp .... ———

ASSAY DATA

M-3, M-4 — Round Number
0.11 | 0.13 — Assay in % Sn

Figure 7. Underground geological plan of main east-west drift (line 6000N) across East Kemptville Tin deposit (after Shell Canada Resources Ltd. Map 763, 1981).

Table 2. Paragenetic sequence, East Kemptville Tin Deposit (after Richardson, 1988a)



rarely in the deposit, are typically found near the granite-metawacke contact (Fig. 5). They have gradational contacts with the enclosing leucomonzogranite. The segregations could be analogous to miarolitic cavities, but are infilled by the constituent minerals.

Type-1 segregations contain pink perthitic microcline, white albite, bluish quartz, minor muscovite and pyrite and do not have alteration envelopes. White albite crystals project inward into Type-2 segregations. Quartz, subhedral zoned cassiterite, molybdenite rosettes, pyrite and green muscovite occur in the central portions of Type-2 segregations. These segregations lack alteration envelopes, and either cross-cut greisen-bordered zones or are cross-cut, offset and sericitized by greisen-bordered zones. Type-3 segregations consist mainly of pyrite, sphalerite, cassiterite, chalcopyrite, fluorite, minor pyrrhotite, interstitial quartz and green muscovite. They are enveloped by small hematitic halos or greisenized leucomonzogranite. These halos imply that the fluids which crystallized Type-3 segregations were not in equilibrium with the enclosing EK leucomonzogranite.

Massive greisen zones are at least 7 m wide, 50 m long and 15 m deep and are composed of several fine-grained rock-types that are mineralogically and texturally distinct, namely: quartz-topaz rock, quartz-mica greisen and green muscovite greisen. The outer boundaries of these zones are characterized by incomplete greisen that grades into EK leucomonzogranite. Massive greisen zones are mineralogically similar to Type-3 pegmatitic segregations, except the dominant F-bearing mineral is topaz, not fluorite.

Quartz-topaz rock is dark gray and has a dull waxy lustre. It contains quartz, topaz, cassiterite, chalcopyrite, sphalerite, pyrite and minor silver-colored mica and forms the high-grade ore (2-6% Sn). The rock is composed of 1 to 5 mm undulose quartz replaced by a seriate mosaic of fine-grained blocky topaz and quartz. Quartz or topaz subhedra can project into 1 to 4 mm blebs of quartz or anhedral base-metal sulfide minerals that appear to infill open spaces. Cassiterite is either subhedral or granular.

Quartz-mica greisen is light gray, has a vitreous lustre and contains quartz, green- and silver-coloured micas and much less topaz, cassiterite, chalcopyrite and pyrite than quartz-topaz rock. It lacks arsenopyrite and wolframite. This rock-type typically contains 0.5 - 1% Sn. In contrast to quartz-topaz rock, infilling textures are rare in quartz-mica greisen. Most feldspars are partially replaced by muscovite. Quartz-mica greisen surrounds quartz-topaz rock.

Small patches of green muscovite greisen consists of muscovite, cassiterite, quartz, sphalerite and rose-coloured anhedral fluorite, but not topaz. Fine-grained muscovite aggregates replace either all pre-existing minerals in EK leucomonzogranite or the inner zones of greisen-bordered zones. Where alteration is advanced, only remnant quartz or cassiterite remain in a green muscovite groundmass.

Veins and fractures with zoned greisen envelopes emanate from massive greisen zones and cross-cut EK leucomonzogranite, Type-2 pegmatitic segregations and felsite dikes. These zones form an orthogonal, en echelon network oriented subparallel (030°/90-70°E) and perpendicular (120°/90°) to the main contact (Richardson, 1983a). Those of the NE-SW trending set are about 1 m apart; those of the SE-NW set are about 2 m apart. Elliptical concentrations of topaz and metallic minerals occur at the intersections. Neither set is preferentially enriched in metallic minerals. Flat-lying greisen sheets such as those described from the Anchor, Australia (Groves and Taylor, 1973) and Cinovec CSSR-DGR (Baumann, 1970) do not occur at East Kemptville.

Greisen-bordered zones can exceed 20 m long and 15 m deep and terminate gradually as the width of the core zone and envelope diminishes. Total width varies from 0.2-50 cm. A greisen-bordered zone consists of a central microfracture or vein (maximum width 2.5 cm) which is enclosed by symmetrically zoned alteration envelopes that average 4 cm on one side. The core zone varies in composition horizontally and vertically over short distances, so crosscutting relationships do not indicate a consistent relative chronology. Symmetrical zones in the greisen envelope from the core zone outward contain various proportions of quartz, topaz and muscovite. Cassiterite and base-metal sulfide minerals occur in the core zone and all zones that contain topaz. Wolframite and arsenopyrite are found in topaz-bearing core zones.

Greisen Minerals

Cassiterite occurs as large (0.5-3 cm), subhedral, color-zoned and twinned crystals in bluish quartz veins and green muscovite greisen. Cassiterite associated with topaz in massive greisen zones and envelopes of greisen-bordered zones is typically disseminated, granular, rarely twinned and shows patchy zoning. Less commonly, cassiterite occurs as aggregates up to 1 cm in size. In pegmatite segregations, cassiterite occurs as twinned and zoned subhedra. Cassiterite contains minor tantalite and chalcopyrite inclusions. Discrete free-standing crystals are rare. Color variation corresponds to trace-element impurities (Ti, Fe, Nb, W and Ta total 3 wt. % maximum; Richardson, 1988a). Fractures in cassiterite are filled by base-metal sulfide minerals or mica. Some grains are rimmed by pyrite.

Topaz forms either fine-grained aggregates that replace large quartz grains or medium-grained blocky euhedra. The euhedra project into anhedral aggregates of base-metal sulfide minerals that appear to have infilled open cavities. The percentage of F in the (F,OH) site is high and varies slightly (88-100%) (Richardson, 1988a). Preliminary examination of inclusions in topaz indicate that several generations of fluid inclusions are present, but no melt inclusions. The earliest-formed fluid inclusions contain a small vapor bubble and several daughter minerals including halite, possible sylvite, a rod-like or platey non- birefringent mineral and a birefringent mineral. The ratio of vapor-to-liquid is constant, suggesting that this fluid was not trapped during boiling.

Sphalerite, chalcopyrite, pyrrhotite and pyrite fill what could be open spaces in quartz-topaz rock and fractures in cassiterite, molybdenite, wolframite and arsenopyrite. Minor stannite rims sphalerite grains and is included in chalcopyrite. Stannite is not associated with cassiterite. Marcasite replaces pyrite and pyrrhotite. Wolframite, arsenopyrite, molybdenite and galena are uncommon. Wolframite and arsenopyrite occur as subhedral crystals and wolframite blades can be enclosed in arsenopyrite. Both minerals are brecciated and infilled by base-metal sulfide minerals. Minor molybdenite and galena occur with base-metal sulfide minerals.

Cassiterite, wolframite, molybdenite and arsenopyrite grains are typically fractured and dismembered. Locally, cassiterite shows anomalous birefringence and strain lamellae parallel to pull-apart edges. In green muscovite greisen, cassiterite is brecciated and grain margins are replaced by mica. These textural features indicate that deformation occurred after the crystallization of the EK leucomonzogranite, quartz-topaz rock and quartz-mica greisen, but before the formation of green muscovite greisen.

Post-Greisen Veins

Each major stage of post-greisen veining is distinguished by mineralogy and cross-cutting relationships (Richardson et al, 1982; Richardson, 1983a; 1988a). These veins contain quartz, sulfide, phosphate, carbonate, sulfate and zeolite (particularly stilbite) minerals, but not cassiterite. Topaz has not been observed, but fluorite is common. Fluorine-bearing minerals such as, fluoro- apatite, triplite (Mandarino et al, 1984), phosphophyllite, childrenite and mcauslanite (Richardson et al, 1988d), are also present. Some veins are composed of ribs of apatite, quartz and/or albite, indicating precipitation occurred during extensional deformation.

Deposit Geochemistry

Representative analyses of the East Kemptville leuco-monzogranite and various greisen types in the deposit are given in Tables 3 and 4 respectively. The geochemistry of the deposit as a whole (for mass balance enrichment/depletion studies) is given by 43 one hundred tonne bulk samples through the main drift zone (Fig. 7) shown in Table 1. Compared to other rock types of the DLC the East Kemptville deposit is significantly enriched in Sn-Cu-Zn-Ag (recovered),As, Sb, Bi, Cd, Mn, Fe, F, Cl, P, S, Li, Rb, Cs, Ta, Nb and U, and severely depleted in REE, Y, Mg, Ti , Th and Zr (see Table 1). Greisen zones in the deposit are highly enriched in Sn, F, Cu, Zn and W but contain variable amounts of Li and Rb depending upon their petrology. The greisens are notably depleted in Na whereas the host leucogranite is much more enriched in this element compared to other rock units in the DLC (see Tables 1, 3 and 4); overall, however, the deposit, as a whole, is depleted in Na and enriched in Cl (Table 1).

Sulfur Isotope Studies

A sulfur isotope study of sulfide phases from the tin and base metal stages of mineralization at the East Kemptville tin-base metal deposit indicates the following features (Kontak, 1990a);

1. The range of $\delta^{34}S$ isotopic values for 31 sulfides analyzed is small with minimum and maximum values of +3.6 and +6.6 per mil.

2. The $\delta^{34}S$ values (0/00) for individual sulfide phases are: pyrite = 5.1 $\pm$ 0.2 (n=14), pyrrhotite = 5.5 $\pm$ 0.1 (n=3), sphalerite = 5.4 $\pm$ 0.3 (n=7), chalcopyrite = 4.9 $\pm$ 0.4 (n=3), arsenopyrite = 6.5 $\pm$ 0.0 (n=3) and galena = 3.6 (n=1).

3. There is no systematic variation of $\delta^{34}S$ minerals for the first 3 stages of mineralization. For example, $\delta^{34}S$ values of pyrite from the tin, sulfide and phosphate stages of mineralization are similar at 4.8 $\pm$ 0.2, 5.1 $\pm$ 0.2 and 5.2 $\pm$ 0.1 per mil, respectively. These data indicate a homogeneous source for the S, absence of fractionation during mineralization and an open system throughout vein formation.

Table 3. Representive analyses of the leucogranite phase in the East Keamptville Tin Deposit (* denotes analysis of trace element by intrumental neutron activation technique; A/CNK = molecular ratio (Al_2O_3/CaO + Na_2O + K^2O); ND = not detected; %Co = % normative corundum calculated on F-free basis), (after Kontak, 1990b in press).

Sample	003	009	012	013A	021	023A	023C	071	072	091	105	109	110	124B	156	161	178	1075B
SiO2	73.16	73.11	75.02	74.14	71.51	74.06	72.32	72.83	72.85	74.00	74.29	73.36	71.85	73.48	72.10	72.97	72.70	73.50
TiO_2	0.00	0.08	0.08	0.08	0.00	0.04	0.00	0.00	0.00	0.00	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.02
Al_2O_3	15.10	14.52	13.88	14.42	15.33	14.89	14.58	14.94	13.51	15.00	14.28	14.55	17.09	15.81	16.00	15.09	14.75	15.70
Fe_2O_3	1.15	1.16	1.05	1.17	0.70	1.38	1.31	1.37	1.45	1.23	1.56	1.29	1.42	1.92	1.17	1.23	1.08	1.32
MnO	0.06	0.04	0.03	0.04	0.15	0.03	0.04	0.06	0.06	0.04	0.08	0.04	0.04	0.04	0.24	0.10	0.04	0.05
MgO	0.03	0.03	0.03	0.04	0.03	0.01	0.03	0.03	0.05	0.03	0.02	0.02	0.02	0.01	0.02	0.03	0.01	0.02
CaO	0.64	0.46	0.40	0.52	0.70	0.49	0.44	0.46	0.66	0.66	1.05	1.12	0.58	0.42	0.25	0.31	0.54	0.49
Na_2O	3.59	3.81	3.90	3.79	5.31	3.77	3.68	3.64	3.50	4.01	3.37	3.89	3.76	3.62	4.12	3.73	4.38	3.88
K_2O	4.08	3.84	3.81	3.96	2.41	4.03	3.93	3.78	3.45	3.50	2.96	4.14	3.83	3.69	3.55	4.10	3.79	3.86
P_2O_5	0.60	0.49	0.53	0.48	0.97	0.43	0.49	0.45	0.51	0.60	0.31	0.75	0.41	0.13	0.46	0.39	0.64	0.45
LOI	0.83	0.71	0.79	0.87	0.62	1.20	0.76	0.81	1.03	0.61	1.14	0.51	0.80	0.68	0.55	0.36	0.52	0.35
	0.65	1.15	0.77	2.01	1.23	0.48	0.69	1.08	1.23	1.19	0.67	0.82	0.75	1.05	0.43	0.48	0.42	0.46
	99.89	99.40	99.52	99.51	98.96	100.82	97.58	99.45	98.30	100.87	99.72	100.50	100.56	00.92	99.24	98.76	99.23	100.86
F=O	0.27	0.48	0.32	0.84	0.51	0.20	0.28	0.45	0.51	0.49	0.28	0.34	0.31	0.44	0.18	0.20	0.17	0.19
	99.62	98.92	99.20	98.67	98.45	100.62	97.30	99.00	97.79	100.38	99.44	100.16	100.25	100.48	99.06	98.56	99.06	100.67
B	<10	<15	<10	<10	<10	<10	<10	<10	<10	<10	-	-	-	-	-	-	-	-
Li	170	234	224	167	218	680	417	754	545	424	454	525	728	970	325	661	670	500
Rb	844	892	894	876	447	929	963	905	838	786	714	814	895	1150	748	991	858	822
Sr	60	33	55	77	450	64	77	39	55	54	221	140	72	33	70	15	52	33
Ba	ND	ND	ND	ND	126	9	ND	24	28	25	45	67	14	9	34	8	27	17
Ba*	-	-	-	28.5	81.4	-	-	-	5.0	-	-	-	14.1	-	21.4	-	-	-
Pb	20	30	45	33	57	66	34	105	99	32	40	24	38	53	67	37	56	59
Ga	32	33	35	36	42	32	38	38	37	38	30	31	33	31	28	31	31	31
Nb	32	33	37	34	63	28	31	26	29	33	30	30	29	26	34	26	30	27
Nb*	-	-	-	13.5	37.7	-	-	-	11.9	-	-	-	16.8	-	25.5	-	-	-
Zr	28	28	30	28	28	34	30	28	28	27	32	32	31	28	29	27	30	33
Zr*	-	-	-	16.1	19.5	-	-	-	12.9	-	-	-	19.7	-	16.7	-	-	-
Y*	-	-	-	4.6	1.9	-	-	-	5.0	-	-	-	4.9	-	3.7	-	-	-
U	21	19	27	16	21	30	27	32	28	29	26	25	28	27	25	27	25	23
U*	17.1	-	-	14.1	-	26.7	24.8	-	26.8	25.9	-	-	-	-	-	-	-	-
Th	5	5	6	5	4	6	5	6	6	6	5	5	6	6	6	5	5	6
Th*	4.5	-	-	4.8	2.8	4.9	5.3	-	5.2	5.5	4.8	-	5.2	-	4.8	-	-	5.5
Sn	77	110	87	101	90	210	97	354	360	73	107	81	114	96	102	NA	230	155
W	14	21	25	11	16	42	24	44	37	13	17	16	18	39	12	23	19	12
W*	11.6	-	-	9.8	-	32	23	-	32	11.4	-	-	-	-	-	-	-	-
Zn	177	43	22	27	373	222	286	112	226	89	307	178	55	170	41	52	331	66
Cu	65	13	7	7	11	86	113	149	155	86	23	82	254	56	13	30	55	23
Mo	4	4	2	<2	27	4	7	3	<2	<2	2	5	2	2	3	2	<2	2
V	ND	ND	ND	ND	3	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	ND	1	ND
Ni	1	2	3	2	2	4	4	8	7	5	3	5	6	2	4	4	6	6
Co*	0.3	-	-	0.3	-	0.4	0.4		0.4	0.3	0.2	-	-	-	0.2	-	-	0.1
Ta	14	10	15	13	36	11	11	11	10	15	14	12	14	13	14	16	13	NA
Ta*	12.6	-	-	11.2	31.9	10.5	11.3	-	7.5	14.1	12.5	-	10.7	-	12.0	-	-	11.2
Hf	<2	<2	<2	<2	3	2	<2	<2	<2	<2	<2	<2	2	<2	2	<2	<2	NA
Hf*	2.1	0.3	-	1.7	2.4	1.9	1.9	-	1.4	2.1	1.9	-	1.7	-	1.8	-	-	2.1
Cs	33	39	49	25	11	42	39	51	32	26	30	29	36	58	26	48	31	32
Cs*	22	-	-	38	-	36	35	-	29	24	-	-	-	-	-	-	-	-
Sc	2.1	1.8	3.9	2.3	0.5	2.2	1.9	2.2	1.8	1.9	2.1	2.6	2.1	2.0	2.3	2.2	2.1	1.9
Sc*	2.2	-	-	2.5	-	2.1	2.3	-	2.2	2.5	2.3	-	-	-	2.7	-	-	2.3
A/CNK	1.31	1.29	1.23	1.26	1.22	1.30	1.31	1.37	1.26	1.29	1.34	1.13	1.50	1.48	1.44	1.35	1.20	1.34
% Co	4.84	4.19	3.37	3.94	4.08	4.37	4.44	4.97	4.18	4.65	4.46	2.88	6.75	5.45	5.47	4.60	3.50	4.45
Na_2O/K_2O	0.87	0.99	1.02	0.95	2.20	0.93	0.93	0.96	1.01	1.14	1.13	0.93	0.98	0.98	1.16	0.90	1.15	1.00

Table 4. Representative analyses of the main types of greisen mineralization in the East Kemptville Tin Deposit (after Richardson et al., 1990a).

Lithology	EK25 QTGSN	EK26 QTGSN	EK31 QTGSN	EK32 QTGSN	EK33 QTGSN	EK34 QTGSN	EK35 QTGSN	EK36 QTGSN	EK40 QTGSN	EK24 MGSN	EK42 MGSN
SiO_2	70.3	67.0	77.9	71.5	73.9	68.8	67.4	75.3	72.0	41.2	38.8
TiO_2	0.04	0.03	0.04	0.04	0.04	0.04	0.03	0.05	0.04	0.07	0.07
Al_2O_3	15.3	14.8	12.3	13.9	12.2	14.8	14.3	14.9	14.3	25.8	24.6
Fe_2O_3	3.75	1.11	3.81	0.83	1.54	3.50	0.84	1.50	2.09	5.50	0.78
FeO	0.8	5.4	0.8	2.6	1.9	0.9	2.3	1.6	1.2	2.3	3.7
MnO	0.01	0.03	0.01	0.15	0.09	0.02	0.07	0.06	0.06	0.45	0.43
MgO	0.1	0.1	0.0	0.1	0.1	0.1	0.1	0.1	0.1	0.2	0.2
CaO	0.21	0.78	0.40	0.53	0.32	0.43	0.97	0.28	0.74	4.65	5.66
Na_2O	0.1	0.1	0.1	0.2	0.2	0.3	0.5	1.0	0.2	0.2	0.2
K_2O	0.16	0.41	0.28	3.30	2.20	0.23	1.70	2.08	1.31	9.52	9.00
P_2O_5	0.34	0.43	0.25	0.38	0.25	0.34	0.49	0.24	0.38	3.03	3.38
LOI	3.85	2.16	3.08	2.31	2.31	4.00	3.08	2.31	3.00	3.54	4.00
%S	3.9	3.4	3.7	1.1	2.0	4.1	2.5	1.6	2.4	0.0	0.0
%F	2.0	6.5	1.6	1.4	1.3	4.4	3.8	1.6	3.4	2.8	1.6
Total	100.9	102.3	104.3	98.2	98.3	102.0	98.0	102.5	101.1	99.2	92.5
Sn	5800	5800	366	3400	3660	3900	5630	527	6100	16900	36800
W	68	82	<10	10	<10	<10	120	40	70	83	28
Mo	2	1	2	6	1	2	2	4	2	5	1
Ca	<1	16	4	30	20	3	23	22	9	49	44
Li	34	180	73	855	405	77	690	815	225	1220	1210
Rb	52	153	113	1279	743	89	737	842	486	2490	2334
Sr	8	26	11	35	22	22	39	77	20	261	295
Ba	18	23	90	120	150	120	140	150	210		120
Zr	22	17	10	<10	<10	<10	<10	10	20	66	50
Y	2	4	<10	<10	<10	<10	<10	<10	<10	43	<10
U	17	10	28	15	43	18	19	20	29	38	51
Th	3	6	9	16	14	14	14	18	20	57	28
Nb	20	20	40	<10	20	30	20	30	40	20	20
Cu	2400	3900	2860	3280	2000	2760	2950	510	7090	22	29
Pb	8	20	20	23	20	20	17	14	20	110	157
Zn	2100	3300	900	2964	7425	12400	17850	604	4839	190	162
Ta	2.6	3.3								13.6	
As	4	10	7	19	7	11	<5	<5	13	15	36
Bi	15	45	22	53	33	39	31	<2	45	2	21
B	<25	<25	1.7							50	
Cl	50	50								250	
Ag			6.3	6.7	4.8	5.0	5.6	0.8	19.3		1.4
La	3.9	5.7									
Ce	8.6	13.1									
Sm	2.3	3.9									
Hf	1.6	1.5									
Eu	0.04	0.32									
Tb	0.0	0.5									
Yb	0.1	0.2									
Lu	0.0	0.0									
c	16.90	15.30	12.22	10.67	10.20	15.82	12.56	11.37	13.21	14.04	14.13
K_2O/Na_2O	1.6	4.1	3.1	18.3	9.6	0.9	3.8	2.0	8.7	50.1	39.1
Al_2O_3/CNK	32.6	11.5	16.0	3.5	4.4	15.9	4.6	4.4	6.5	1.6	1.7
Rb/Sr	7	6	10	37	34	4	19	11	2	10	8
Eu/Eu*	0.03	0.22									
Ca/Yb	86.0	68.9									

Oxides; S, F in weight percent; others in parts per million. QTGSN = quartz-topaz rock and quartz-mica greisen; MGSN = green muscovite greisen.

4. The $\delta^{34}S$ mineral values are interpreted to indicate that the dominant form of S in the ore fluid was H_2S. Calculations of $\delta^{34}S$ fluid using mineral-H_2S fractionation factors and temperatures derived from fluid inclusion studies indicate a primary $\delta^{34}S$ fluid value of 5.0 $\pm$ 0.5 per mil, consistent with a magmatic source for the S.

5. Consistent values of $\delta^{34}S$ for sulfides from the main stages of mineralization indicate that incursions of extraneous S were either nonexistent or minimal. This is consistent with O isotopic data (Kontak, 1988), but contrasts with the Sr isotopic results of Richardson et al (1989) and Richardson et al. (1990b) which indicate, therefore, that multiple fluids may have been involved in vein formation.

6. Comparison to known high ^{34}S reservoirs in the Meguma Terrane are consistent with a dominantly magmatic source of the S since interaction with Meguma Group sedimentary rocks is known to shift $\delta^{34}S$ values to more positive values (e.g., peribatholithicmineralization and Meguma gold deposits).

7. The $\delta^{34}S$ values for sulfides from East Kemptville are characterized by their narrow range and enrichment in heavy S compared to many of the more significant centres of Sn and W mineralization. These features reflect a homogeneous source of S which was itself enriched in ^{34}S compared to source regions for S in, for example, the Panasqueira, Pasto Bueno and Chojlla deposits.

KEMPT SNARE LAKE CUPOLA (KSLC)

A small leucogranite - leucomonzogranite cupola hosting W-Pb-Zn-As-Cu-Ag mineralization is present at Kempt Snare Lake (KSLC) along the southern extension of the postulated East Kemptville Shear Zone (Fig. 3; Soehl, 1988; Soehl et al, 1988). The KSLC has been variably greisenized and displays a progressive alteration sequence similar to the Davis Lake Complex (Table 1) which progresses with increase of secondary muscovite from leucogranite to greisenized granite, black megacrystic greisen and fine grained black greisen (Fig. 8). Biotite-bearing phases such as those occurring in the Davis Lake Complex are not present in this cupola. Soehl et al, (1988) note the presence of a coarse grained leucomonzogranite in fault contact with the leucogranite (drill indicated) which they suggest may be a less differentiated granite emplaced from deeper levels by faulting.

The cupola is characterized by both ductile and brittle deformational features which intensify with degree of hydrothermal alteration.

Mineralization within the KSLC is restricted mainly to quartz veins and their immediate wallrocks. Arsenopyrite, sphalerite, argentiferous galena, scheelite and chalcopyrite with minor siderite and fluorite make up the economic minerals within the quartz vein swarms.

Progressive alteration in the KSLC is characterized by an increase of K_2O, Al_2O_3, H_2O+, Rb and C and decrease of SiO_2, Na_2O, CaO and Sr. With the exception of K_2O and C the geochemical pattern for alteration follows that of the Davis Lake Complex. The KSLC is notable for its high graphite content compared to other greisen deposits in the Meguma Terrane. Compared to the muscovite-bearing and greisenized phases of the Davis Lake Complex the KSLC rocks are much more potassic, anomalously low in Sn, Cu, Zn, and P, and enriched in Th (see Tables 1 and 5).

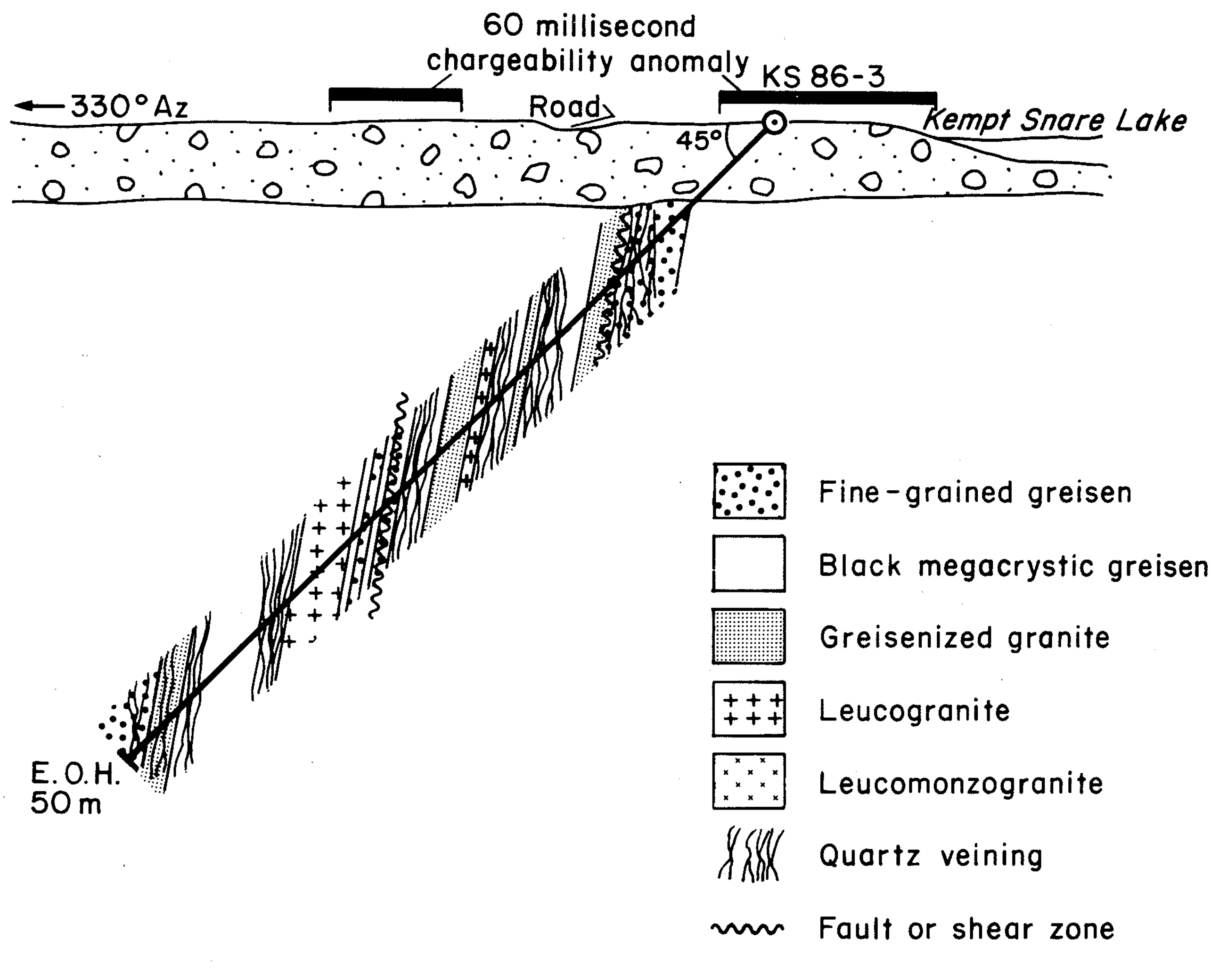


Figure 8. Generalized geological drill log section for Kempt Snare Lake prospect (scale 1" = 10m; after Soehl et al., 1987).

Table 5. Representative geochemical analyses of the various granitic rocks and alteration facies of the Kempt Snare Lake cupola (oxides in wt. %; all other values in ppm except where noted; after Soehl et al., 1987).

SAMPLE	86-2-8	86-1-2	86-3-3	86-1-4	86-1-6	86-1-1	86-2-4	86-3-1	86-2-3	86-3-9
ROCK TYPE[I]	LMG	LGT	LGT	GG	BMG	BMG	BMG	BMG	FGG	FGG
SiO_2	76.69	75.81	76.17	75.64	70.02	75.20	74.75	73.31	74.34	75.43
TiO_2	0.10	0.07	0.06	0.04	0.09	0.06	0.05	0.07	0.07	0.04
Al_2O_3	12.46	12.88	12.89	13.59	15.70	13.50	13.51	14.59	13.59	13.39
Fe_2O_3	0.43	0.36	0.35	0.36	0.69	0.30	0.45	0.79	0.74	0.44
FeO	0.60	0.67	0.60	0.47	1.27	0.57	0.70	0.87	0.70	0.70
MnO	0.09	0.05	0.05	0.02	0.12	0.07	0.14	0.07	0.03	0.04
MgO	0.69	0.72	0.61	0.71	0.80	0.78	0.72	0.76	0.83	0.73
CaO	0.49	0.49	0.32	0.32	0.06	0.39	0.19	0.11	0.05	0.31
K_2O	4.55	4.50	4.24	4.50	6.63	5.40	5.50	6.06	5.86	4.54
Na_2O	2.95	3.50	3.57	3.62	1.42	2.91	2.84	2.03	2.33	3.39
$H2O^+$	0.62	0.62	0.60	0.66	1.44	0.85	0.94	1.04	0.89	0.72
$H2O^-$	0.08	0.10	0.11	0.07	0.10	0.10	0.07	0.18	0.10	0.10
P_2O_5	0.08	0.08	0.09	0.09	0.09	0.09	0.08	0.08	0.08	0.09
CO_2	0.71	0.28	0.40	0.14	0.34	0.32	0.37	0.81	0.08	0.39
C	0.03	0.03	0.02	0.03	0.71	0.09	0.22	0.53	0.44	0.07
TOTAL	100.49	100.07	99.97	100.18	99.38	100.53	100.46	101.12	100.03	100.28
Fe_2O_3 TOT	1.10	1.11	1.02	0.88	2.10	0.93	1.23	1.76	1.52	1.22
TRACES (ppm)										
As	18	124	275	63	1020	158	210	200	331	58.6
Ba	60	74	78	54	99	87	58	69	73	23
Bi	0.04	7.30	0.04	0.04	0.04	0.04	0.04	0.04	0.07	0.05
Cu	4	1	1	1	1	1	6	1	1	1
F	4080	1963	1380	820	2020	950	1260	1200	1600	2350
Mo	2	2	2	2	3	2	2	2	1	6
Pb	41	16	22	3	6	18	20	27	43	26
Rb	522	540	465	518	901	603	626	778	695	625
Sn	5.7	5.7	4.9	5.8	16.0	8.2	8.6	12.0	8.5	5.2
Sr	43	61	31	48	14	29	21	15	16	20
Ga	19	22	21	23	31	22	22	27	23	26
Zn	60	30	142	28	436	50	94	95	124	45
U	26.4	28.5	25.9	31.8	30.0	23.9	28.8	27.3	30.2	33.7
W	11	6	6	22	25	13	10	19	13	22
B	14	12	11	12	55	32	37	55	28	12
Y	60	44	47	40	75	57	55	66	43	44
Li	133	125	135	129	228	92	147	226	199	154
Th	25	24	15	16	20	19	24	25	23	16
Ta	3.2	3.3	5.0	5.2	4.2	6.7	3.0	4.6	3.8	5.5
Hf	5	3	3	3	4	3	4	5	3	3
La	17	7	9	4	2	10	8	6	3	5
Nb	12	11	13	12	16	14	10	13	13	14
Cs	10	13	8.3	11	13	10	9.4	12	11	12
Ce	42	24	28	16	10	27	23	19	14	19
Zr	76	63	55	55	79	66	66	65	67	49
Ag	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
Au(ppb)	<2	<2	<2	<2	<4	<2	<2	<2	<2	<2

LMG = Leucomonzogranite GG = Greisenized Granite FGG = Fine Grained Greisen
LGT = Leucogranite BMG = Black Megacrystic Greisen

Age dating of muscovites and K-feldspars in this deposit by $^{40}Ar/^{39}Ar$ stepwise outgassing method show minimum overprinting events of 265 Ma and 220 Ma for muscovite and K-feldspar respectively. For highly altered samples a maximum plateau age of 300 Ma was obtained while less altered samples yielded a maximum plateau apparent age of 330 Ma. The latter age is considered to be only a minimum age for intrusion of the pluton due to the widespread presence of deformation and alteration features (Soehl et al, 1988).

WEDGEPORT PLUTON

The Wedgeport Pluton is a small intrusive body at the southern extremity of the East Kemptville Tin Domain (Fig. 1). It has been dated by Cormier et al (1987) at 316 Ma (U-Pb on zircon) making it the youngest of the intrusive bodies in the Tin Domain. The pluton is characterized by a main body of biotite monzogranite with lesser amounts of biotite-muscovite monzogranite, albitized granite, greisenized granite and greisen (Chatterjee et al, 1985). The biotite monzogranite shows textural varieties from equigranular to porphyritic.

The following description of the greisen mineralization in this pluton is given by Chatterjee et al (1985):

"Greisenized granite occurs usually around the greisen bodies and is characterized by marked hydrothermal alteration dominated by muscovitization and chloritization. Biotite typically forms euhedral to subhedral books scattered evenly through the greisenized granite. The biotite is generally accompanied by a group of minerals which are mainly interstitial or occur with interstitial minerals, namely fluorite, tourmaline, apatite, muscovite, topaz, cassiterite and chlorite, and it is devoid of inclusions of zircon and apatite. Muscovite and chlorite occur as thin rims on biotite, in zones cutting biotite crystals and also as flakes intergrown with biotite. Accessory minerals consist of tourmaline, sphalerite, chalcopyrite, arsenopyrite, cassiterite, wolframite, scheelite, native bismuth, argentite, pyrite and pyrrhotite.

The greisen zones range in width from a few centimetres to over three meters, and compositionally are extremely varied. Some are dominated by muscovite while others are essentially quartz veins. The sulphide content of the greisens range from less than five percent to more than 70 percent. The major gangue minerals of the greisens are quartz and phyllosilicates together with topaz, fluorite, beryl, tourmaline and carbonates. The common ore minerals are the same as those observed in the greisenized granite."

Primary fluid inclusions in fluorite and quartz in the greisen veins of the Wedgeport Pluton have moderate salinities and homogenize at ca. 295°C (Al et al, 1986).

GENETIC MODELS OF GRANITE-HOSTED DEPOSITS

Theories on the genesis of individual mineral deposits (e.g. East Kemptville Tin deposit) and the East Kemptville Polymetallic Tin Domain are still evolving. Richardson (1990) proposes the following model for fluid evolution in the formation of the East Kemptville deposit.

The primary chemical specialization of the DLC magma and the subsequent vertical zonation through the magma chamber culminated, at the present erosional level, in the crystallization of the EK leucomonzogranite. During the last stages of crystallization in the deposit, over 75,000 tonnes of F separated from this residual magma and entered a previously exsolved aqueous phase (Richardson et al., 1990a). Pegmatitic segregations found near the granite-metawacke contact reflect the sequential evolution of this "transitional" (Burnham and Ohmoto, 1980) magmato-hydrothermal aqueous fluid. Early metal- and F-poor pegmatitic segregations lack alteration envelopes, indicating that the fluid was not corrosive at this time. Envelopes around later metal- and F-rich segregations are similar to those that enclose massive greisen and greisen-bordered zones (Richardson et al., 1988a).

Field relationships indicate that significant amounts of Rb-rich, Sr-poor EK leucomonzogranite interacted with the F-rich greisen-forming fluid. The alkali and alkaline earth elements released from EK leucomonzogranite were not accommodated in the minerals characteristic of East Kemptville greisen (topaz, quartz, cassiterite and base-metal sulfides). Mica, a possible host, is limited in the massive greisen. Comparison of the chemical composition of EK leucomonzogranite and quartz-topaz rock indicated that the greisen-forming fluid contained an extremely large quantity of F and probably also had high Rb, K, Na, Cs and Li contents and low Sr contents. Consequently it had high Rb/Sr. Such an unusual, highly corrosive fluid is not analogous to basinal brines that form sediment-hosted Cu, Pb or Zn deposits or Mississippi Valley type deposits (Jackson and Beals, 1967; review by Anderson and Macqueen, 1987), or the saline brines generated in porphyry deposits (Titley and Beane, 1981). A closer analog is the dense, but usually miscible, alkali-,volatile- and silicate-rich fluids that are associated with rare-metal pegmatites (London, 1987).

In the deposit, the ore-forming greisen fluid was reduced by the precipitation of cassiterite and made less corrosive by the precipitation of large quantities of F-rich topaz. The residual post-greisen fluid was then incapable of replacing leuco-monzogranite, so fluid pathways within the deposit became spatially restricted. The occurrence of phosphates and base-metal sulfide minerals, stilbite (Na-bearing zeolite) and dickite in the post- greisen veins and on joint surfaces reflects the still substantial metal Cu, Fe, Zn, S, P, Na and K contents of the fluid residual to the formation of the cassiterite-topaz ore.

Although the metawacke above the contact is largely eroded, preserved roof pendants provide an opportunity to evaluate the degree of interaction between the Meguma Group metawacke and the greisen fluid. Veins are present that contain minerals typical of the greisen assemblage (cassiterite, muscovite, arsenopyrite, other sulphide minerals, beryl, tourmaline), but these veins extend only a limited distance (<50 m, and typically several meters) into metawacke. These veins have much smaller alteration envelopes (1-3 cm) compared to the greisen-bordered veins in the granite (30-50 cm). Skarn is not present at the deposit, but locally, the chloritized metawacke is replaced by white mica.

The absence of a highly fractured carapace and lack of pressure-quench textures indicate that hydrostatic pressure did not repetitively exceed lithostatic pressure during crystallization and ore formation (Richardson et al., 1988a). The East Kemptville deposit is thus unlike porphyry-style Cu-Mo or Mo-W deposits. The relatively impermeable and chemically unreactive nature of the overlying metawacke helped retain Rb- and F-rich ore-forming fluid within the EK leucomonzogranite.

The presence of incomplete greisen, quartz-mica greisen and the gradationally zoned greisen envelopes that surround some veins, fractures and massive greisen zones indicate that in the deposit, much of the ore fluid was buffered by the EK leucomonzogranite. The rare occurrences of cassiterite, arsenopyrite and pyrrhotite in the topaz-rich cores of large greisen-bordered zones and massive greisen zones suggests the greisen fluid was internally buffered only in these areas (cf. Heinrich and Eadington, 1986). In contrast, the veins which contain the greisen-assemblage minerals in the overlying metawacke are enveloped by thin alteration envelopes. These veins can have well-defined walls with inward-facing muscovite sprays. These observations suggest that, in the metawacke, the greisen was internally buffered (cf. Heinrich and Eadington, 1986).

$^{87}Sr/^{86}Sr$ systematic studies suggest that the greisen fluid was modified during the waning stages of mineralization by fluids from the overlying Meguma Group metawacke (Model 2; Richardson et al, 1990b). This Meguma-modified fluid was retained in and near the deposit by small-scale recirculating convection cells. With time and decreasing temperature, the isotopic composition of this fluid decreased to bulk-rock metawacke values. Rb and Sr contents also lessened.

With regard to the origin of the East Kemptville Leucogranite hosting tin mineralization at the East Kemptville Tin deposit Kontak (1990b) presents the following model:

A petrological study of the EKL, a muscovite-topaz leucogranite hosting tin mineralization, reveals the following important features:

(1) The granite represents the crystallization product of highly evolved, volatile-rich (i.e., F) felsic melt. The uniform textures and model mineralogy and absence of extreme alkali metasomatism are interpreted to indicate magmatic origin for both topaz and muscovite rather than products of late-stage metasomatism.

(2) There is an absence of petrographic and textural features (e.g., graphic textures, miarolitic cavities, breccias, abundant pegmatites, unidirectional solidification textures (USTs)) which are commonly found in fluid-saturated melts (e.g., Kirkham & Sinclair, 1988) indicating, therefore, that the EKL did not reach fluid saturation (cf. Richardson et al., 1988b).

(3) Mineralogically and chemically the EKL is similar to other fluorine-rich felsic sites including topaz granites. It is characterized by elevated contents of Cs, F, Li, Nb, Rb and Ta, and depletion in B, Ba, Ca, Co, Hf, Mg, Mn, Ni, Sc, Ti and V.

(4) The strongly peraluminous character (A/CNK values = 1.13 to 1.84) and major and trace element chemistry of the EKL, including strongly depleted REE contents (particularly Eu), indicate that the granite represents the product of fractional crystallization rather than batch melting. Although it is not presently possible to unequivocally eliminate the DLC as a parental magma to the EKL, the field relationships as currently understood suggest this may be valid. Alternatively, magmatic evolution may have proceeded at depth with no manifestation of this process at surface.

(5) The chemistry of the EKL, including whole-rock O isotopic data, indicates a crustal reservoir for the protolith. Analogies with other F-rich suites suggest that the most likely source is a felsic granulite, probably a residue from earlier melting processes. Hence, the EKL is interpreted to reflect the incongruent breakdown of a fluorine-rich biotite phase due to crustal anatexis related to injection of basaltic magmatism to the lower crust. The source rock is considered to be represented by similar lithologies exposed in the Liscomb Complex (Giles & Chatterjee, 1987).

Boyle (1990) gives the following explanation on the origins of the East Kemptville Polymetallic Tin Domain:

Radiometric, gravity and geochemical data indicate that the Davis Lake Pluton is probably a distinct granitic body intrusive into the much larger South Mountain Batholith to the north and the Meguma Terrane to the east and southwest.

Greisenization processes in the Davis Lake Pluton are characterized by progressive hydrothermal alteration of a parental biotite monzogranite to form biot-musc monzogranite, musc-monzogranite, greisenized monozogranite and greisen stockwork. This sequence displays, on a mass balance adjusted basis (Gresen's Formula), marked progressive increases in Sn, Cu, Zn, Cd, As, Sb, Bi, Ag, Li, Rb, Cs, F, Cl, P, Fe, Mn, Mo, W, Nb, Ta, Co, Ni, Be, Tl, U, Sr, Ca, Al and CO_2; no discernable changes in Si, V, Pb, Hg, and Se; and marked progressive decreases in K, Na, Mg, Ti, Th, Sc, B, Zr, Ba, Y, and La.

The East Kemptville Tin province is characterized by passive ascent of a magmatic felsic melt initially undergoing orthomagmatic processes resulting in formation of a chilled fractured contact carapace and followed by a retrograde boiling H_2O-saturated melt/aqueous/volatile process characterized by generation of a H_2O-H_2S/SO_2-HF/SiFx-HCL/Cl-PO_4/HPO_4 aqueous phase enriched in Sn, Zn, Cu, W, Li, Rb, As, and Mn. Greisenization of the carapace zone and mineralization of the overlying Meguma Terrane is accomplished by both mass transport and diffusive movement of these fluids through the fractured carapace into favourable structural zones (cusps, cupolas, contact inflections, permeable transition meta-sediments). Volume changes in the retrograde boiling zone related to

H_2O-saturated melt > crystal > aqueous/volatile processes, together with high wallrock rigidity and lithostatic confining pressures, are the probable driving and constraining forces for this system.

Later, post-Visean, tectonic dislocation of the magma/ aqueous/volatile system gave rise to rapid vapor/aqueous release of $HCl-H_2O$ (F poor) fluids to form highly silicified kaolin deposits in large shear and breccia zones.

MEGUMA-HOSTED TIN DEPOSITS

Introduction

Tin mineralization in the Meguma metasediments comprises a number of vein and stratiform deposits in which cassiterite is characteristically associated with chloritization (Fig. 1, 3 and 4). The three main occurrences as discussed below are the Duck Pond, Pearl Lake and Dominique(Plymouth) prospects. The various Meguma- hosted deposits are very similar in their structural settings and types of mineralization but differ in their geochemical and petrological characteristics. A comparison of the geochemistry of the Duck Pond and Pearl Lake deposits together with data for background host metawackes and meta-argillites is presented in Table 6. The Duck Pond deposit, besides being enriched in Sn, is also enriched in many of the elements (F, Cu, Zn, Mn, Ag, Mo, As) concentrated in the granite-hosted East Kemptville deposit, suggesting a genetic link between these two deposits. The Pearl Lake deposit is much more enriched in F, Cu, and As than the Duck Pond deposit but the latter contains much higher concentrations of Zn, Pb, Mn, Mo and Ag.

Table 6. Geochemical composition of Duck Pond and Pearl Lake tin mineralization and unmineralized metasediments in the northern and southern granite-Meguma contact zones (oxides in weight percent; all other elements in ppm except where noted; after Boyle unpublished data).

	Mineralization in Metasediments				Unmineralized Metasediments					
					Northern Contact				Southern Contact	
	Duck Pond Meta-argillite		Pearl Lake Metawacke		Meta-argillite		Metawacke		Metawacke	
	n=15		n=23		n=12		n=15		n=36	
	AM(S.D.)	GM	AM(S.D.)	GM	AM(S.D.)	GM	AM(S.D.)	GM	AM(S.D.)	GM
SiO_2	56.15 (4.67)	56.23	72.21 (6.92)	72.44	62.76 (7.37)	63.10	74.85 (4.88)	74.13	73.83 (6.03)	74.13
TiO_2	.83 (.06)	.83	.60 (6.92)	.58	.80 (.13)	.79	.56 (.11)	.55	.54 (.15)	.51
Al_2O_3	20.57 (1.77)	20.42	12.17 (2.97)	11.75	17.63 (4.18)	16.98	12.18 (2.16)	12.02	12.87 (2.60)	12.59
Fe_2O_3T	10.11 (4.08)	9.77	5.55 (3.03)	5.13	3.88 (2.51)	6.17	3.79 (1.07)	3.63	3.82 (1.51)	3.55
CaO	.11 (.07)	.09	.77 (.28)	.71	.65 (.37)	.53	1.10 (.31)	1.05	.87 (.39)	.79
MgO	2.26 (.46)	2.24	1.49 (.11)	1.41	2.27 (.50)	2.23	1.18 (.41)	1.12	1.25 (.59)	1.12
Na_2O	.76 (.99)	.55	2.22 (.88)	1.86	2.70 (2.59)	1.78	2.69 (.45)	2.63	2.15 (.63)	2.04
K_2O	3.40 (.79)	3.24	2.26 (.79)	2.15	3.14 (1.35)	2.75	2.07 (.85)	1.91	2.81 (.92)	2.69
P_2O_5	.10 (.02)	.10	.12 (.03)	.12	.17 (.04)	.16	.12 (.04)	.11	.13 (.03)	.11
CO_2	.12 (.06)	.11	.10 (.0)	.10	15 (.04)	.11	.13 (.03)	.11	.11 (.03)	.11
S	1161 (2246)	288	3238 (5590)	832	157 (279)	85	105 (113)	65	51 (35)	42
F	790 (187)	758	1259 (549)	1148	680 (239)	616	477 (162)	457	459 (190)	426
Cl	122 (41)	115	256 (79)	245	290 (270)	214	245 (124)	213	246 (121)	229
B	42 (18)	38	38 (9)	36	25 (10)	20	20 (4)	19	55 (34)	34
Li	57 (16)	56	52 (16)	50	52 (28)	43	32 (14)	28	35 (20)	30
Rb	189 (77)	155	148 (51)	138	142 (54)	129	82 (35)	76	135 (48)	126
Cs	7 (5)	6	7 (4)	6	7 (4)	5	7 (4)	6	6 (3)	5
Sr	161 (211)	105	152 (68)	129	170 (68)	151	199 (56)	190	144 (54)	135
Ba	688 (182)	646	509 (127)	490	616 (293)	524	480 (178)	389	515 (223)	467
Sn	629 (865)	246	525 (712)	178	10 (6)	7	10 (6)	7	8 (6)	5
W	3 (2)	3	4 (4)	3	3 (3)	2	5 (6)	3	4 (3)	3
Nb	19 (4)	18	13 (5)	12	15 (3)	14	9 (2)	9	10 (3)	10
Ta	2 (.5)	2	1 (.2)	1	2 (.8)	2	2 (.4)	2	2 (.3)	2
V	106 (30)	83	54 (25)	41	60 (32)	49	53 (22)	47	54 (21)	50
Zr	125 (23)	123	188 (60)	160	224 (106)	209	217 (42)	214	197 (40)	195
U	2.5 (.3)	2.5	2.0 (.6)	1.9	2.6 (.8)	2.5	1.9 (.4)	1.8	3.7 (2.7)	3.2
Th	16.6 (7.4)	15.5	12.2 (8.9)	8.5	11.3 (4.6)	10.2	8.7 (2.7)	7.9	9.9 (3.0)	9.5
Cu	36 (39)	25	151 (247)	55	12 (12)	8	16 (13)	12	19 (11)	16
Pb	57 (71)	37	7 (12)	4	5 (4)	3	8 (6)	6	11 (6)	9
Zn	1190 (1900)	603	542 (1567)	162	62 (15)	60	76 (31)	71	48 (15)	46
Mn	4148 (2364)	3715	931 (1051)	708	917 (381)	851	543 (239)	501	502 (179)	478
Ag	.76 (1.9)	.23	.22 (.45)	.10	.14 (.19)	.09	.26 (.39)	.13	.07 (.06)	.06
Cd	11.7 (29.5)	2.4	4.4 (12.9)	.8	.12 (.1)	.10	.17 (.10)	.10	.13 (.14)	.10
Mo	2.3 (1.9)	1.7	1.1 (1.2)	.8	1.1 (.7)	.9	1.0 (.8)	1.0	1.2 (.6)	1.0
Ni	35 (5)	35	17 (7)	16	27 (7)	27	30 (10)	29	25 (8)	23
Co	14 (6)	12	9 (9)	7	8 (3)	8	12 (7)	11	10 (4)	9
Hg(ppb)	20 (18)	16	11 (3)	11	14 (10)	12	20 (15)	16	38 (22)	30
As	22 (17)	15	130 (55)	75	12 (13)	6	7 (4)	6	5 (6)	3
Sb	.4 (.2)	.4	.8 (1.2)	.4	.5 (.4)	.4	.7 (.5)	.6	.3 (.1)	.3
Se	.3 (.1)	.3	.3 (.15)	.3	.3 (.1)	.3	.3 (.0)	.3	.3 (.0)	.3

Duck Pond Prospect

The Duck Pond prospect occurs in metawacke and meta-argillite of the transition unit between the Goldenville metawacke and the Halifax meta-argillite. When compared to the Goldenville Formation metawacke, the transition unit metawacke contains finer-grained quartz and more sericite (15 - 60%). Sericite separates the quartz grains and preclude quartz-quartz sutured boundaries. The quartz grains also disrupt the planer fabric of the meta-argillite and increase the initial permeability of the metawacke. Transition-unit metawacke contains less Si, K, Na, Ca and Sr and more Al when compared to Goldenville metawacke in the Davis Lake area and elsewhere in Nova Scotia.

Meta-argillite of the transition unit consists of sericite, accessory chlorite and visible lensoid aggregates (0.4 mm) of rutile. Amorphous iron oxyhydroxides are disseminated throughout the rock-type. Cavities 1 - 4 mm in size are present in the meta-argillite and could represent dissolved concretions. This unit contains less silica, more alumina and alkali elements when compared to the transition unit metawacke.

In the Duck Pond area, the transition unit dips approximately 70°NW. Although the regional metamorphic grade is the chlorite zone of greenschist facies, garnet occurs as disseminated metamorphic porphyroblasts, in calcareous concretions and in veins. The narrow post-orogenic contact metamorphic aureole of the Davis Lake complex does not extend to the Duck Pond prospect.

Quartz veins that contain minor hematite pseudomorphs of pyrite, and trace sphalerite and chalcopyrite occur throughout the transition unit. Unlike most veins in this unit, these veins do not have alteration envelopes. In the meta-argillite, the hematite-bearing quartz veins follow the slaty and crenulation cleavages. In the metawacke, these veins are poorly defined with irregular margins. As this vein set is corrugated and cross-cut by all other veins, it is likely related to folding that predated the intrusion of the Davis Lake complex.

Vein Mineralization and Alteration

Alkalic, chloritic or argillic alteration zones envelope the veins that comprise the Duck Pond prospect (Table 7, Fig. 9). Cassiterite is associated with chloritic alteration. In the meta-argillite, metallic minerals occur in well-defined veins that are typically parallel to the slaty and crenulation cleavages. In the metawacke, metallic mineral zones are diffuse, follow originally more permeable layers or microfractures and reflect the strata-bound nature of this type of mineralization. Later processes affecting minerals in all vein groups include recrystallization, stabilization of marcasite, sulphidation of chalcopyrite to bornite and covellite, and the oxidation of pyrite.

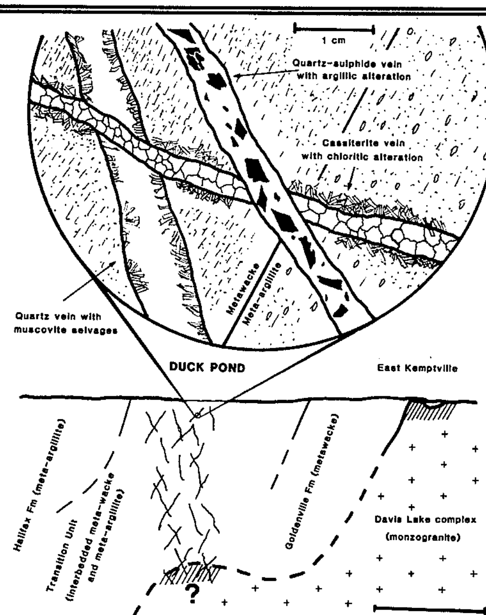


Figure 9. Idealized northwest looking cross-section of the Duck Pond tin prospect. Shape of granite sedimetn contact from Egner (1987). Enlargement: schematic representation of the vein paragenesis (after Pitre and Richardson, 1989).

Table 7. Paragentic sequence, Duck Pond prospect (after Pitre and Richardson, 1989).

Time (not to scale) →

425°C — Temperature (approximate) → 200°C

Mineral	Associated alteration: Alkalic	Chloritic		Argillic		Oxidation
Anorthoclase	**					
Muscovite	******					
Quartz	*****	****	********	*****	**	
Cassiterite						
I		****				
II			**			
III			*			
IV			****			
Wolframite		*				
Garnet		****	**			
Chlorite		****	*******	*	*	
Tourmaline			**			
Chalcopyrite	*	**	**		**	
Sphalerite	*	**	**		**	
Arsenopyrite		*		*****		
Pyrrhotite			**		**	
Pyrite				*****[a]		
Chalcostibite					**	
Bornite					*	
Covellite					*	
Marcasite					*	
Hematite						**
Fracturing Period	I	II	III	IV	V	

Alkalic alteration

Alkalic alteration is characterized by muscovite-anorthoclase veins and quartz veins with muscovite selvages. It is associated with Period 1 fracturing (Table 7). Anorthoclase forms equant grains (0.25 mm) disseminated throughout thin (0.4 mm) muscovite veinlets. In the wider (5 mm) muscovite-quartz veins, muscovite is restricted to the vein wall. Here, the muscovite forms a fine-grained and unoriented layer which provides a base for coarser-grained (1 mm) radiating muscovite that projects into the vein. Minor fine-grained chalcopyrite and sphalerite form interstitial inclusions in quartz. Euhedral recrystallized clusters of pyrite occur along fractures.

Chloritic alteration

At Duck Pond, several different habits of cassiterite are associated with chloritic alteration. In stratabound layers up to 1 m thick in metawacke, cassiterite is disseminated as concentrations that reach 2% by volume in a matrix of quartz and chlorite and/or garnet. In meta-argillite, cassiterite forms discrete anastomosing veins of >99% cassiterite with chlorite selvages. In both rock-types, cassiterite occurs as coarse-grained yellow sub- to euhedral crystals (1 mm, cassiterite I), and less commonly as red equant euhedral crystals (0.4 mm, cassiterite II), or orange coliform layers (cassiterite III). Tan, fine-grained, acicular radiating aggregates (0.05 mm, cassiterite IV) are also present. Duck Pond cassiterite is not zoned or twinned, unlike cassiterite at East Kemptville.

Chloritized meta-argillite consists of massive chlorite and rutile. In this rock-type, cassiterite I forms discrete irregular veins up to 1 cm wide along Period II fractures. Cassiterite I is massive and contains chlorite, muscovite, anorthoclase, quartz, chalcopyrite and sphalerite inclusions. Chalcopyrite and sphalerite inclusions are absent at the vein walls, and increase towards the interior of the vein. Trace bornite, covellite and pyrite rim cassiterite or occur along microfractures.

Well-indurated chloritized metawacke is mottled with interpenetrating lath-shaped patches of chlorite (0.2 x 1.0 cm). These areas consist of fine-grained chlorite (0.25 mm long) that crosscut strained quartz grain boundaries. The habit of the chlorite suggests that it pseudomorphs medium-grained mica. In this rock-type, cassiterite I is disseminated and contains inclusions of chalcopyrite, sphalerite and quartz. Cassiterite II containing chalcopyrite and pyrrhotite inclusions is concentrated along Period III fractures. Minor tourmaline (pleochroic: blue, colourless, brown) and trace chalcopyrite, sphalerite and pyrrhotite are spatially related to cassiterite II. Cassiterite III fills cavities in metawacke. Cassiterite IV is disseminated and forms radiating aggregates in metawacke and meta-argillite that is altered to chlorite.

Garnet is ubiquitous in chloritized metawacke and its concentration is typically proportional to the sericite:quartz ratio. Cassiterite I and II, tourmaline, chalcopyrite, sphalerite and pyrrhotite are randomly included in garnet. Pyrite euhedra up to 5 mm contain chalcopyrite inclusions. Garnet and massive cassiterite-rich regions exhibit quartz-filled fractures related to Period IV fracturing.

Argillic alteration

Argillic alteration is the most common alteration type at Duck Pond. In meta-argillite, symmetrical argillite alteration halos around the veins are typically about 1 cm wide. The contact between the alteration zone and the unaltered host rock is sharp. The mineralogy of the altered meta-argillite is similar to the unaltered host rock except the altered meta-argillite does not contain disseminated chlorite nor iron oxyhydroxides, but does contain minor euhedral pyrite (0.5 mm) with chalcopyrite, sphalerite and pyrrhotite inclusions. In metawacke, argillic alteration has created anastomosing translucent dull grey planar zones that are leached of chlorite and iron oxides. These zones are centred on microfractures or follow the more permeable cross- beds. The minerals are not symmetrically distributed within the alteration zones in the metawacke.

In the meta-argillite, veins with argillic alteration are symmetrically zoned inward. A thin (0.5-2 mm) quartz layer is in contact with the wall-rock. Rare cassiterite and brecciated wolframite are found with trace arsenopyrite, chalcopyrite and sphalerite in this layer. Orange garnet subhedra (0.4 mm) form a layer above the quartz and the crystal faces are developed inwards towards the centre of the vein. Cassiterite I, quartz, arsenopyrite, chalcopyrite, sphalerite and pyrrhotite are randomly included in garnet. The interior part of these veins contain discrete grains of quartz, pyrite, arsenopyrite, minor chalcopyrite, sphalerite, bornite, covellite and trace chalcostibite ($Cu_6Tl_2SbS_4$). Pyrrhotite occurs as inclusions in pyrite, arsenopyrite, sphalerite and chalcopyrite. Pyrite and arsenopyrite occur as euhedral crystals. Pyrite also includes quartz, garnet, chalcopyrite and sphalerite. Discrete chalcopyrite grains are rimmed by bornite and covellite. Covellite also occurs along microfractures in quartz, cassiterite, chalcopyrite and sphalerite.

Arsenopyrite occurs as either optically continuous "islands" in quartz or as fractured grains that have poorly-fitted edges on fragment margins. The interstices between the arsenopyrite fragments can be filled with quartz, chlorite, chalcopyrite, sphalerite or chalcostibite. These unusual replacement textures appear to indicate that arsenopyrite was dissolved and replaced by quartz. If so, according to the calculations of Heinrich and Eadington (1985), this processes must have occurred as the oxygen fugacity rose.

Later processes

The base-metal sulphide minerals formed during alkalic, chloritic and argillic alteration were affected by sulphidation, recrystallization, temperature-related mineral inversion and oxidation. Sulphidation reactions include pyrrhotite to pyrite, chalcopyrite to bornite and covellite, and bornite to covellite. Pyrrhotite is typically found as inclusions in other minerals. This relationship and the paucity of pyrrhotite in the Duck Pond prospect suggest that pyrrhotite exposed along mineral grain boundaries was converted to pyrite. Bornite is uncommon at Duck Pond, but where present, it rims chalcopyrite. Covellite rims chalcopyrite, sphalerite and bornite. Covellite also occurs in microfractures and as inclusions in sphalerite.

Although arsenopyrite was brecciated during Period V fracturing, pyrite neither fills the interstices between fragments nor is brecciated. This suggests that pyrite precipitated before, and was recrystallized after, arsenopyrite was brecciated. Radial vermicular strained quartz often surrounds euhedral pyrite suggesting such recrystallization did occur. Also, chalcopyrite, sphalerite and pyrrhotite always have mutual boundaries and are present as rounded inclusions in pyrite - textures that are suggestive of recrystallization. Arsenopyrite and chalcostibite are relatively "hard" minerals and are less susceptible to recrystallization relative to other base-metal sulphide minerals (Skinner et al. 1972).

Pyrite is sometimes replaced by marcasite, but pyrrhotite is not. In contrast, at East Kemptville, both pyrite and pyrrhotite are replaced by marcasite. At Duck Pond, hematite replaces pyrite and cross-cuts argillic alteration.

Genesis of Duck Pond Deposit

Cassiterite at Duck Pond is found in four different habits in veins found in meta-argillite and disseminated in metawacke; stannite is not present. The alteration styles, their crosscutting relationships and the fracture-filling habit of the metallic minerals suggest that the Duck Pond veins formed by repeated hydrostatic fracturing due to volatile release from the underlying Davis Lake complex while it was crystallizing.

Much of the cassiterite was precipitated under conditions externally buffered by iron oxyhydroxides in the wall rock. During argillic alteration, arsenopyrite and pyrite were precipitated and sulphidation of pyrrhotite to pyrite is suspected. Pyrrhotite, pyrite, chalcopyrite and sphalerite were subsequently recrystallized. Sulphidation of chalcopyrite yielded bornite and covellite. Marcasite was stabilized. Using mineral assemblages and textural relationships, it appears that as the temperature dropped from 425-405°C to 200°C at Duck Pond, the pH ranged from 5.2 to no lower than 3 and log fO_2 dropped from at least -19 to -43. Log aS_2 rose from <-15 to >-10. Cassiterite precipitated at higher temperature and pH and lower log aS_2.

Pearl Lake Prospect

Mineralization at the Pearl Lake Prospect (Fig. 4) is very similar to that of Duck Pond, with the exception that the former is considerably enriched in F, Cu and As, and contains less Zn, Pb, Mn, Mo, and Ag (see Table 6). The vein mineralogy at Pearl Lake consists of quartz, carbonates, fluorite, garnet, sphene, chlorite and cassiterite (Chatterjee, 1979). As at Duck Pond cassiterite is mainly associated with chloritization.

Dominique (Plymouth) Prospect

Tin mineralization at the Dominique prospect and other smaller tin occurrences in the Meguma metasediments of the Plymouth area (see Fig. 1) have been classified by Wolfson (1983) into three main types, namely: A). sulphide-cassiterite veinlets in metasediments with restricted chlorite alteration. B). stratiform sulphide-cassiterite replacement bodies in calcareous layers, and C). rare detrital cassiterite grains in a pelitic microscour. Types A and B, occurring along the northern contact of the Wedgeport Pluton (Fig. 1), are presumed to be associated with mineralizing events (mainly tin) in this intrusion. With the exception of type C, mineralization in this area is similar to that at Duck Pond and Pearl Lake. The pre-tin, tin and post-tin stages of vein mineralization noted at Duck Pond (Pitre and Richardson, 1990) are also present at the Plymouth occurrences (Wolfson, 1983). A notable difference between these two areas with regard to tin-stage mineralization is the presence of carbonates (calcite, siderite) and scheelite (wolframite at Duck Pond) in the Plymouth deposits.

The Plymouth Meguma-hosted deposits display an outward zonation pattern from the Wedgeport Pluton characterized by Mo-W enrichment within 2 km of the pluton, Sn mineralization 3-4 km away and finally Pb-Zn 4-7 km distant.

Road Log

Due to the lack of outcrop in the East Kemptville area the field trip to these deposits will be confined to an examination of the open pit at the East Kemptville (Sn, Cu, Zn) deposit and examination of 'type' drill core sections from the area. In addition a tour of the 'state of the art' mill at the East Kemptville mine has been arranged.

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